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Abstract

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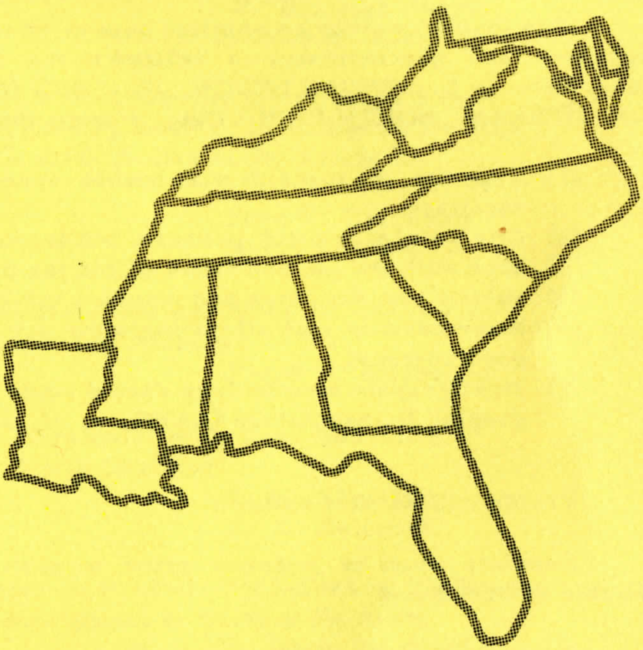

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SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 20, No. 1

November 1978

STRATIGRAPHIC REVISION OF THE EXPOSED MIDDLE EOCENE TO LOWER MIOCENE FORMATIONS OF NORTH CAROLINA[1]

By

Gerald R. Baum
Department of Geology
College of Charleston
Charleston, South Carolina 29401

W. Burleigh Harris
and
Victor A. Zullo
Department of Earth Sciences
University of North Carolina at Wilmington
Wilmington, North Carolina 28403

ABSTRACT

Tertiary strata between the middle Miocene Pungo River Formation and the Paleocene Beaufort Formation, which have been generally grouped and mapped as the Eocene Castle Hayne Limestone, are divisible into five distinct litho- and chronostratigraphic units: the middle Eocene Castle Hayne Limestone, the upper Eocene New Bern Formation, the Oligocene Trent Formation, and the lower Miocene Belgrade and Silverdale Formations. These formations are delimited in outcrop by the Cape Fear arch to the southwest and the Neuse River fault to the northeast.

The middle Eocene Castle Hayne Limestone consists dominantly of two laterally equivalent lithofacies: bryozoan biosparrudite and bryozoan biomicrudite. A basal phosphate-pebble biomicrudite is locally developed. The New Bern Formation, with the type section designated the Martin Marietta quarry at New Bern, disconformably overlies the Castle Hayne Limestone. The principal lithology is a sandy, pelecypod-mold biomicrosparrudite. No diagnostic microfauna or megafauna occur in the New Bern Formation; therefore, placement of the unit in the upper Eocene is based principally on stratigraphic position. The Oligocene Trent Formation disconformably overlies the New Bern Formation. The modified type section is along the Trent River from New Bern to within 0. 7 km of Pollocksville, The Trent Formation consists of three

[1] Contribution Number 736 to Marine Sciences at the University of North Carolina at Wilmington.

lithologies: a basal sandy, echinoid biosparite; a sandy, pelecypod-mold biomicrudite, and an upper barnacle, pelecypod-mold biosparrudite. Diagnostic fauna include Solidobalanus (Hesperibalanus) n. sp., Pecten aff. P. perplanus poulsoni and Chlamys trentensis (Harris).

Strata exposed at Belgrade and Silverdale, traditionally considered part of the Trent Formation, do not correlate with the type section of the Trent Formation. They lie disconformably on the Trent Formation and are designated the Belgrade Formation and the Silverdale Formation of early Miocene age. The Belgrade Formation, type section at the Martin Marietta quarry at Belgrade, consists of a lower quartz arenite facies that grades upward into a sandy, pelecypod-mold biomicrudite facies. Molluscan holotypes from the type section were described by Richards (1948). The laterally equivalent Silverdale Formation, type section at the quarry located just east of the town of Silverdale, represents a downdip, offshore facies of the Belgrade Formation. The principal lithology of the Silverdale Formation is an unconsolidated, sandy, pelecypod biomicrudite. Molluscan holotypes from the type section were described by Kellum (1926, locality 10 655). Barnacles of the Balanus concavus stock characterize both formations.

INTRODUCTION

Purpose

The most concerted research efforts on the Eocene Coastal Plain stratigraphy of North Carolina have generally been at quarry sites located at the northeastern and southwestern limits of the outcrop belt (New Bern; Castle Hayne-Wilmington). The combined effects of restricted observations, dearth of natural outcrops and poor fossil preservation have led to a general misunderstanding and misinterpretation of North Carolina Tertiary stratigraphy. Therefore, the main purpose of this study is to expand the area of investigation to include the intervening and outlying regions, to differentiate and delineate mappable time-stratigraphic units and to suggest lateral and vertical facies relationships. The emphasis is to provide a reasonable and practical interpretation of the lithostratigraphic framework of the Tertiary formations that crop out between the middle Miocene Pungo River Formation and the Paleocene Beaufort Formation.

Regional Setting

The Atlantic Coastal Plain Province consists of an oceanward thickening wedge of Mesozoic-Cenozoic sediments and rocks. In North Carolina, this homoclinal relationship is locally disrupted by three prominent structural features: the Cape Fear arch, the Cape Lookout-Neuse River fault zone (Neuse fault in this report), and a northeast-southwest trending lineament herein referred to as the Carolina fault (Figure 1).

The Cape Fear arch, for years recognized as the major structural feature of the Atlantic Coastal Plain Province (Dall and Harris, 1892), has also been referred to as the Great Carolina ridge and the Wilmington anticline. Stephenson (1926) mapped and delineated the axial trace of the structure. Subsequent work by MacCarthy (1936), Mansfield (1937), Richards (1945), and Straley and Richards (1950) have confirmed the presence of a structural high that parallels the Cape Fear River. The Cape Fear arch plunges to the southeast at 2.5-2.8 m/km (Spangler, 1950; Bonini and Woollard, 1960) and is asymmetrical with the steeper flank to the northeast (Maher, 1971).

Hersey et al. (1955) indicated that the Blake Plateau is an offshore expression of the Cape Fear arch; however, some authors have considered the Blake Plateau to be a depositional feature (Heezen et al., 1966; Pratt, 1968). Bonini and Woollard (1960) suggested that movement on the arch occurred twice: once in the Late Cretaceous and once in the Late Tertiary. Harris (1975; 1976; 1978) has since documented movement during the Maestrichtian, and Harris et al. (1977) and Baum (1977) have documented movement during the middle Eocene. In addition, Zullo and Harris (1978) have suggested movement during the Pleistocene.

The identification of a northwest-southeast positive element trending parallel to the Neuse River was based on isopachous mapping and structural contouring of the Yorktown Formation (Gibson, 1967; 1971); however, Ferenczi (1959) considered the positive element a fault and designated it the Cape Lookout-Neuse River fault zone (the Neuse fault).

LeGrand (1955) and Ferenczi (1959) postulated another possible fault zone trending northeast-southwest which passes through the vicinity of Kinston, Lenoir County. The unnamed fault, herein designated the Carolina fault, was suggested by the occurrence of salt water incursion near the confluence of the Cape Fear and Black rivers (LeGrand, 1955). The Carolina fault and the Neuse fault have apparently been periodically active throughout the Tertiary (Baum et al., 1977; Baum, 1977).

In North Carolina, outcrops of Eocene to lower Miocene strata are bounded by two northwest-southeast trending structurally positive features: the Cape Fear arch to the southwest and the Neuse fault to the northeast. These two structural features delineate and confine the area of investigation to the counties of Craven, Jones, Onslow, Pender, New Hanover and Duplin (Figure 1). Outcrops of younger sediments are essentially confined to the area northeast of the Neuse fault; whereas south along the Cape Fear arch, post-depositional erosion has removed most of the Tertiary section.

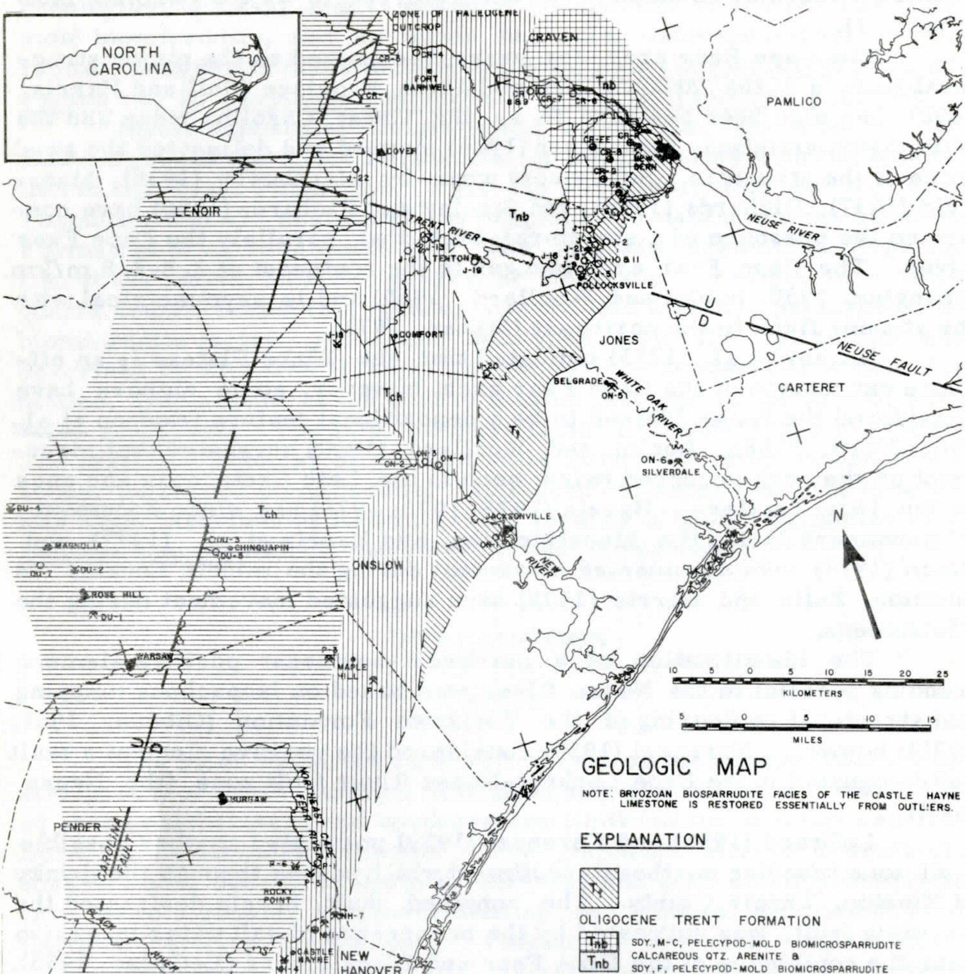

Figure 1. Geologic map of middle Eocene to lower Miocene strata of southeastern North Carolina. (Note: The areal distribution of the Castle Hayne Limestone, biosparrudite facies, is reconstructed from outliers to represent its occurrence at the time of deposition).

Acknowledgments

Martin Marietta Aggregates has been more than generous in allowing us to work in their quarries and allowing access to their 2-inch exploratory cores. We are also indebted to Geotechnical Engineering Company for giving us 2-inch cores drilled in Brunswick and New Hanover Counties. Also, we thank independent operators throughout eastern North Carolina for permitting unlimited access and study of their quarries.

STRATIGRAPHIC REVISIONS

"Strata of Two Ages"

The general misunderstanding and misinterpretation of North Carolina Tertiary Coastal Plain stratigraphy was perpetrated early in Coastal Plain geologic studies by the confusion of Cretaceous and Eocene strata in the Wilmington-Castle Hayne area of southeastern North Carolina. Although Lyell (1845) was the first to recognize Eocene strata in North Carolina at an exposure near Wilmington, the exact placement of the Mesozoic-Cenozoic boundary eluded American geologists (e. g. Hodge, 1841; Miller, 1912; Brown et al., 1972). Stanton (1891) was the first to recognize the exact nature of the Mesozoic-Cenozoic boundary and placed it above a "siliceous limestone bed" and below a basal conglomerate (phosphate pebble biomicrudite) of the Castle Hayne Limestone.

Fallaw and Wheeler (1963) and Wheeler and Curran (1974) presented paleontological data that confirmed the placement of the boundary immediately below the basal conglomerate of the Castle Hayne Limestone. Wheeler and Curran (1974) formally named the Cretaceous carbonate the Rocky Point Member of the Peedee Formation (Figure 2). Nevertheless, some workers (Bain, 1970; Brown et al., 1972) still consider the carbonate in New Hanover County which occurs below the basal phosphate pebble biomicrudite of the Castle Hayne Limestone to be part of the middle Eocene.

Harris and Bottino (1974) and Harris (1976) dated glauconite pellets from the Rocky Point Member using Rb/Sr ratios. An average model age of 67. 6 m. y. and an isochron age of 68. 1 m.y. also confirmed that the Rocky Point Member was Late Cretaceous. Harris (1975; 1976; 1978) has since mapped the subsurface distribution of the Rocky Point Member.

Castle Hayne Limestone (Restricted)

Type Section. The Castle Hayne Limestone was named by Miller (1912) for "exposures in the vicinity of the town of Castle Hayne." The

SERIES \ AUTHOR	MILLER 1912	KELLUM 1926	BROWN 1955, 1958	SWIFT & HERON, 1969: WHEELER & CURRAN, 1974	BROWN, MILLER & SWAIN, 1972 (#FM. NAMES NOT USED BY AUTHORS)	THIS REPORT
LOWER MIOCENE		TRENT FORMATION	ABSENT IN NC		UNNAMED- MAPPED WITH OLIGOCENE	CRASSOSTREA / SILVERDALE FM. / BELGRADE FM
OLIGOCENE			ABSENT IN NC		UNNAMED	TRENT FORMATION
UPPER EOCENE	CASTLE HAYNE LIMESTONE ? CRASSOSTREA ?	CRASSOSTREA / CASTLE HAYNE LIMESTONE	CASTLE HAYNE LS. INCLUDES CRASSOSTREA — UPPER EOCENE COMPONENT		ABSENT IN NC	NEW BERN FORMATION
MIDDLE EOCENE	TRENT FM.- ?IN PART K CARBONATE		CASTLE HAYNE LS. INCLUDES CRASSOSTREA — MIDDLE EOCENE COMPONENT	CASTLE HAYNE LIMESTONE (AGE NOT SPECIFIED)	#NEW BERN FM. / #CASTLE HAYNE LS. / #ROCKY PT. MEM.	CASTLE HAYNE LIMESTONE
UPPER CRETACEOUS	PEEDEE FORMATION	K CARBONATE / PEEDEE FORMATION	PEEDEE FORMATION	ROCKY PT. MEM. / PEEDEE FORMATION	#PEEDEE FORMATION	ROCKY PT. MEM. / PEEDEE FORMATION

Figure 2. Chronological development of portions of the Tertiary stratigraphy of North Carolina.

type section is herein designated as the section exposed in the Martin Marietta quarry on County Road 1002, 3.4 km northeast of the town of Castle Hayne, New Hanover County (Figure 1; locality NH-1-D). At this locality, the Castle Hayne Limestone has a maximum thickness of 11 m (Figure 3) and is characterized by the presence of bryozoans and sponges (Upchurch and Textoris, 1973; Upchurch, 1973). Portions of the section are partially to completely dolomitized (Cunliffe, 1968; Harris et al., 1977; Baum, 1977). The lithology and fauna of the Castle Hayne Limestone in the area of the type section have been described by Canu and Bassler (1920); Kellum (1926); Cooke (1959); Cooper (1959); Kier (1962); Cunliffe (1968); Upchurch and Textoris (1973); Upchurch (1973); and Baum (1977).

Distribution. Outcrops of the Castle Hayne Limestone are essentially confined by the Cape Fear arch to the southwest and the Neuse fault to the northeast (Figure 1). The lower boundary of the Castle Hayne Limestone is marked by disconformities with the Paleocene Beaufort Formation (CR-1), the Cretaceous Peedee Formation (DU-1, DU-5) and the Cretaceous Rocky Point Member of the Peedee Formation (P-3, NH-7, NH-1-D, NH-2, NH-B-2, NH-CB-27). The upper boundary is marked by disconformities with the Yorktown Formation (CR-5), the New Bern Formation (CR-7, CR-8, CR-9) and the Duplin Formation (DU-7). Along the New River, the Oligocene Trent Formation may rest disconformably on the Castle Hayne Limestone.

The Castle Hayne Limestone consists of three facies, in ascending stratigraphic order: phosphate pebble biomicrudite, bryozoan biosparrudite and bryozoan biomicrudite. The phosphate pebble biomicrudite facies forms a discontinuous conglomerate at the base of the Castle Hayne Limestone. It is best developed in New Hanover and Pender counties where it lies disconformably on the Cretaceous Rocky Point Member.

In New Hanover County, the bryozoan biosparrudite facies consists of erosional remnants (Figure 3). It is either entirely absent or is separated from the overlying bryozoan biomicrudite facies by a diastem. In Pender (P-3), Onslow (ON-3) and Craven (CR-5) counties, this facies interfingers with and is gradational with the overlying biomicrudite facies.

The bryozoan biomicrudite facies has been the most intensely studied facies of the Castle Hayne Limestone as the type section consists almost entirely of this facies. Probably all of the Castle Hayne Limestone bryozoans studied by Canu and Bassler (1920), as well as most of the molluscs studied by Kellum (1926), the echinoids by Cooke (1959) and the brachiopods by Cooper (1959) were collected from this facies. Since the bryozoan biosparrudite facies is more easily eroded, the transition from the bryozoan biosparrudite facies to the bryozoan biomicrudite facies approximates the updip limit of the outcropping Castle Hayne Limestone; thus the bryozoan biosparrudite facies is rarely present in outcrop. There is some evidence that the bryozoan

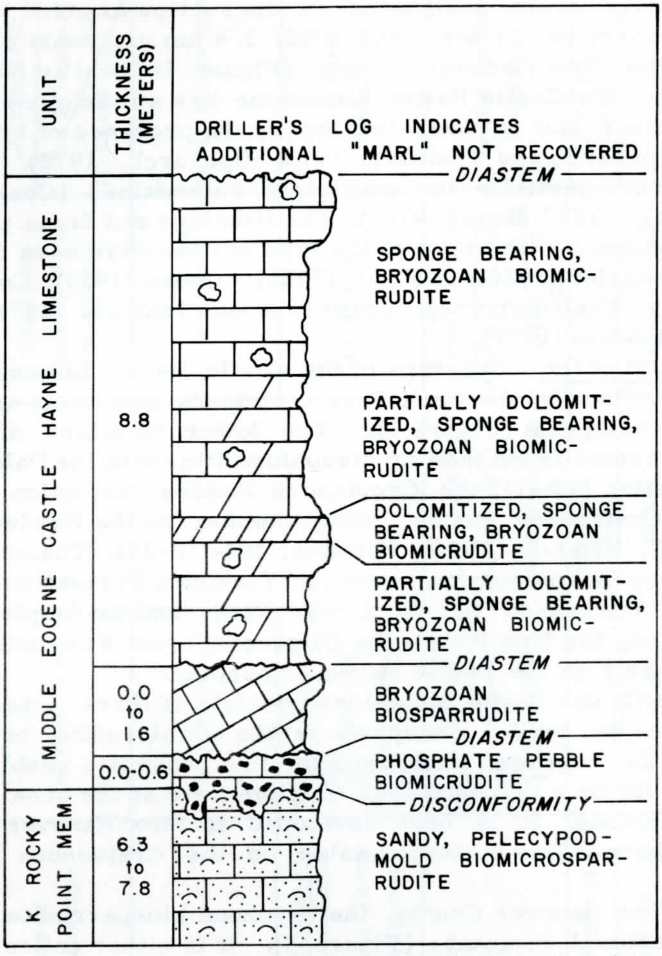

Figure 3. Composite section at the type locality of the Castle Hayne Limestone (NH-1-D).

biomicrudite facies grades basinward into a sandy, foraminiferal biomicrite (CR-7, CR-8, CR-9); however, subsurface work is necessary to establish its distribution.

Correlation. Traditionally, the Castle Hayne Limestone has been considered Jackson in age (Clark, 1909, 1912, Canu and Bassler, 1920; Kellum, 1925; 1926; Cheetham, 1961; Copeland, 1964). Subsurface work by Brown (1958, p. 6) indicated that no lithologic discontinuities exist between Claiborne and Jackson ostracod faunas of the Castle Hayne Limestone and that the Castle Hayne Limestone "was deposited during a temporal transgression from Claiborne time into Jackson time, the bulk of the deposition having occurred during Jackson time."

However, Brown et al. (1972) now consider the Castle Hayne Limestone to be Claiborne in age (Figure 2). Faunal evidence suggests that portions of the Castle Hayne Limestone may extend into the Jackson (Zullo and Baum, in preparation).

New Bern Formation

Type Section. The New Bern Formation is named for a predominantly sandy, pelecypod-mold biomicrosparrudite that disconformably overlies the Castle Hayne Limestone in the Martin Marietta quarry at New Bern, Craven County (locality CR-R). The lithology of this 9.2 m thick section was described by Thayer and Textoris (1972) who recognized three units (which they assigned to the Castle Hayne Limestone) (Figure 4): (1) a basal, 6 m thick, homogeneous and massively bedded, sandy, pelecypod-mold biomicrosparrudite; (2) an overlying sandy, pelecypod-mold biosparite and biosparrudite that ranges from 1.4 to 1.8 m in thickness and grades westward to sand and a calcite-cemented quartz sandstone; and (3) an uppermost, 1.4 m thick, well indurated, massively bedded, sandy, pelecypod-mold biomicrosparrudite. At the type section, the uppermost unit of the New Bern Formation is disconformably overlain by the Oligocene Trent Formation (restricted) or the Yorktown Formation. The contact is marked by phosphatized crusts with attached oysters and barnacles.

Distribution. Unlike the lithologies exposed in the type area of the Castle Hayne Limestone, the type section of the New Bern Formation reveals a thick section of a dominant sandy, pelecypod-mold biomicrosparrudite (see Textoris, 1967; Textoris et al., 1972; Thayer, 1972a; 1972b; Thayer and Textoris, 1972; Baum, 1977).

At three localities (CR-7, CR-8, CR-9), the New Bern Formation overlies the Castle Hayne Limestone. The boundary between the two formations is marked by phosphatized crusts, rounded phosphatized clasts of the underlying unit and a concentration of oysters. The phosphatized crusts have been subsequently bored by pholad pelecypods. The presence of pholad borings indicates that the surface between the two units must have been lithified prior to subsequent deposition of the New Bern Formation. The occurrence of the borings are diagnostic of unconformities (Evans, 1970). Additional evidence suggests that a disconformity exists between the Castle Hayne Limestone (restricted), and the overlying New Bern Formation:

- intertidal, *Callianassa* burrowed and red algae-bearing (*Archaeolithothamnium*) calcareous quartz arenite between the Castle Hayne Limestone and the New Bern Formation indicates a distinct stratigraphic break (Baum, 1977).
- the facies of the New Bern Formation represent a transgressive sequence, rather than a regressive sequence; both the Castle Hayne Limestone and New Bern Formation internally have their own characteristic shallowwater and deeper water facies (Baum, 1977).

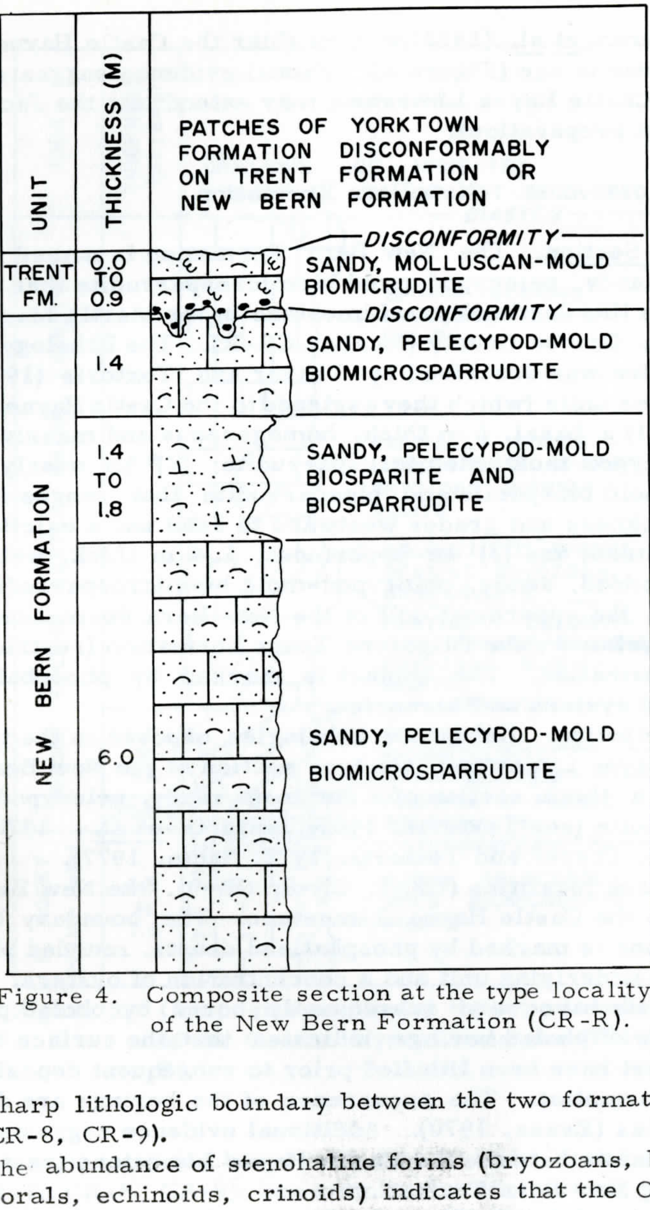

Figure 4. Composite section at the type locality of the New Bern Formation (CR-R).

- sharp lithologic boundary between the two formations (CR-7, CR-8, CR-9).
- the abundance of stenohaline forms (bryozoans, brachiopods, corals, echinoids, crinoids) indicates that the Castle Hayne Limestone was deposited in an open environment with normal marine salinity.
- lack of interfingering of the Castle Hayne Limestone and the New Bern Formation; recognition of the Cretaceous Rocky Point Member of the Peedee Formation no longer necessitates or implies interfingering.
- lack of faunal continuity; the abundant bryozoan fauna characteristic of the Castle Hayne Limestone is absent from the New

Bern Formation - Callista, the dominant pelecypod in the New Bern Formation, as well as, Panopea and Chlamys aff. C. cawcawensis, do not occur in the Castle Hayne Limestone.

- lack of structural continuity; the depositional strike of the Castle Hayne Limestone is approximately northeast-southwest; whereas, the depositional strike of the New Bern Formation is approximately north-south (Figure 1).

Outcrops of the New Bern Formation are confined to an area lying between the Neuse and Trent Rivers (Figure 1) and are essentially limited to central Craven County and central and western Jones County. The New Bern Formation probably does not disconformably overlap any formations older than the Castle Hayne Limestone. The New Bern Formation is disconformably overlain by the Trent Formation and the Yorktown Formation (CR-R, J-8).

The New Bern Formation consists of three facies, in ascending stratigraphic order: calcareous quartz arenite; sandy, fine, pelecypod-mold biomicrosparrudite and sandy, medium to coarse, pelecypod-mold biomicrosparrudite. Along the Neuse River (CR-7, CR-8, CR-9), the New Bern Formation lies disconformably on the Castle Hayne Limestone. At these localities, the calcareous quartz arenite facies can be considered absent. However, this facies appears to thicken along an east-west axis that passes through the vicinity of New Bern.

The sandy, fine, pelecypod-mold biomicrosparrudite facies only crops out at one locality (J-16) where it interfingers with the lower calcareous quartz arenite facies; however, it interfingers with the overlying sandy, medium to coarse, pelecypod-mold biomicrosparrudite facies in several cores (CR-R, CR-J, CR-G, CR-B). Like the calcareous quartz arenite facies, this facies can be considered absent at the Neuse River localities (CR-7, CR-8, CR-9) where the New Bern Formation rests disconformably on the Castle Hayne Limestone. Likewise, this facies appears to thicken along a east-west axis that passes through New Bern.

The sandy, medium to coarse, pelecypod-mold biomicrosparrudite is dominated by generally unabraded pelecypods, mainly of the genus Callista. The fauna also contains Crassetella cf. C. alta, Chlamys aff. C. cawcawensis, Calyptraea, Lucinia, Panopea, Glycymeris and Ostrea. Although mined extensively at New Bern for rock aggregate, this facies has a very limited outcrop. It crops out in Craven (CR-6, CR-7, CR-8, CR-9) and Jones (J-5) counties. It does not appear to crop out south of the Trent River. This facies is frequently confused with the Cretaceous Rocky Point Member of the Peedee Formation.

Correlation. Brown et al. (1972; well CR-T-1) found a diagnostic middle Eocene microfauna within what appears to be the New Bern Formation; however, they also found a diagnostic middle Eocene fauna in the Cretaceous Rocky Point Member (well PEN-OT-9) (Figure 2). Cibicides speciosus, a diagnostic Jackson age foraminifera (considered diagnostic by Brown et al., 1972) was recovered from 2-inch cores (CR-R, CR-G, CR-J). In contrast, the New Bern Formation contains Crassetella cf. C. alta and Chlamys aff. C. cawcawensis, both of which are considered diagnostic of the Claiborne. Of interest, the middle Eocene Castle Hayne Limestone contains Periarchus lyelli, a Jackson zone fossil in the Gulf Coastal Plain. It occurs with Cubitostrea sellaeformis, a diagnostic Claiborne oyster (Stenzel, 1949).

The New Bern Formation does not lend itself well to detailed faunal analyses. Most of the mega-invertebrates are now molds, and the indurated nature of the unit prohibits detailed foraminiferal studies. The most important aspect is that this "facies" is a distinct mappable lithologic unit bounded by disconformities. Thus, this "facies" meets all the criteria of a formation (Ashley et al., 1933). Tentatively, the New Bern Formation is assigned a Jackson or late Eocene age.

Trent Formation (Restricted)

Type Section. The type section of the Trent Formation was designated the Trent River from New Bern to Trenton (Miller, 1912); however, the type section contains two distinct formations. The lower unit belongs to the New Bern Formation, therefore it is proposed that the name Trent be restricted to the upper formation. The modified type section of the Trent Formation is along the Trent River from New Bern to within 0.7 km of Pollocksville. At the type section, the Trent Formation (restricted) lies disconformably on the New Bern Formation.

Distribution. The Trent Formation appears to be confined to an area lying between the Neuse and New rivers (Figure 1). It crops out extensively along the Trent River from New Bern to within 0.7 km of Pollocksville. At New Bern, the Trent Formation is represented by a compressed 1 m section; however, along an axis parallel to the White Oak River, the section thickens and extends updip. Thus the Trent Formation appears to lie within a structural basin delineated by the Neuse and New Rivers.

The Trent Formation lies disconformably on the New Bern Formation at two localities (CR-R, J-8). The upper boundary of the Trent Formation is marked by a disconformity with the Yorktown Formation (CR-R).

The Trent Formation consists of three facies, in ascending stratigraphic order: sandy, echinoid biosparite; sandy, pelecypod-mold biomicrudite and barnacle, pelecypod-mold biosparrudite. All three lithologies are exposed along the Trent River. The fauna of the sandy, echinoid biosparite facies includes the pelecypods Chlamys trentensis, Pecten aff. P. perplanus poulsoni (not P. elixatus) and Anomia; bryozoans, echinoids and the barnacle Solidobalanus (Hesperibalanus) n. sp. which is generally attached to the pectens. This facies lies disconformably on the New Bern Formation at two localities (CR-R, J-8).

The fauna of the sandy, pelecypod-mold biomicrudite is dominated by molluscs which include: Mercenaria, Conus, Panopea, Glycymeris, Calyptracea and vermetid gastropods. Superficially, this facies resembles the sandy, medium-coarse, pelecypod-mold biomicrosparrudite facies of the New Bern Formation; however, it can be differentiated from the New Bern Formation by the presence of Mercenaria, an Oligocene to Recent genus (Cox et al., 1969). At New Bern (CR-R), this facies is represented by a compressed 0.2 m section. At this locality, this facies is characterized by Turritella gastropod molds and a lack of quartz sand. Along the Trent River, this facies appears to thicken and to become more sandy. The increased sand content is due to the fact that the Trent Formation overlaps the New Bern Formation further updip and lower in the section, thus deriving additional sand from the sandier and less consolidated portions of the New Bern Formation. Along the Trent River, this facies interfingers with the lower sandy, echinoid biosparite facies (J-11, J-21) and the overlying barnacle, pelecypod-mold biosparrudite facies (J-1). An outcrop near Jacksonville (ON-1) indicates that this facies had a wide distribution.

The barnacle, pelecypod-mold biosparrudite facies contains opercular, compartmental and basal disc plates of the barnacle Solidobalanus (Hesperibalanus) n. sp. Along the Trent River, this facies thickens and interfingers with the lower sandy, pelecypod-mold biomicrudite facies (J-1).

Correlation. In the Martin Marietta quarry at New Bern (CR-R), the New Bern Formation is overlain disconformably by the Trent Formation, and the disconformity is well documented (Thayer and Textoris, 1972). Approximately 1 m of the Trent Formation is present and contains a new barnacle species of Solidobalanus (Hesperibalanus) that is allied to Eocene and Oligocene species reported from the Gulf and Pacific Coasts of North America, and from western Europe. Along the Trent River, the Trent Formation is exposed from New Bern to within 0.7 km of Pollocksville. At the base of the section, Pecten aff. P. perplanus poulsoni (not P. elixatus) and Chlamys trentensis occur, as well as, the barnacle Solidobalanus (Hesperibalanus) n. sp. Mercenaria, an Oligocene to Recent genus (Cox et al., 1969) is also a common element of the fauna. The concurrence of Mercenaria and Solidobalanus (Hesperibalanus) n. sp. indicates an Oligocene age, and age alluded to by Harris in 1919. The extensive eustatic sea level drop in the mid-Oligocene (Vail, 1976; Gussow, 1976) suggests that this formation is early to middle Oligocene.

Belgrade Formation/Silverdale Formation

Type Sections. The Martin Marietta quarry (ON-5), presently located south of the White Oak River in the town of Belgrade, Onslow County, is designated the type section of the Belgrade Formation. Approximately 6.7 m of section are exposed in the quarry and reveal essentially two facies: 3.0 m of a basal, unconsolidated quartz arenite with abundant Anomia and barnacles which grades up into 3.7 m of a sandy, pelecypod-mold biomicrudite. Locally, a diastem is present in the upper lithology. In places, channel sands containing Crassostrea gigantissima lie disconformably on the Belgrade Formation.

The quarry located at the intersection of County Roads 1434 and 1442, just east of the city of Silverdale, Onslow County (ON-6) is designated the type section of the Silverdale Formation. Two lithologies are present in the quarry: 0.8 m of a basal, dense, sandy, pelecypod-mold biomicrudite which grades upward into 2.4 m of an unconsolidated, sandy, pelecypod biomicrudite. The upper unit is characterized by whole molluscs.

Holotypes of molluscan molds described from the Belgrade Formation include: Modiolus stuckeyi Richards 1948; Panope intermedia Richards 1948; and Cardium belgradensis Richards 1948. The Belgrade Formation is characterized by barnacles of the Balanus (Balanus) concavus stock.

Holotypes of molluscs (as whole shells) described from the Silverdale Formation include all those species described by Kellum (1926) from locality 10655, as well as subsequent additions to the fauna by Richards (1943) from the Gillet and Askew marl pits near Silverdale.

Exposures of the Belgrade and Silverdale Formations have been traditionally confused with the Trent Formation. Also, both quarry sites have been treated by paleontologists as separate entities, principally owing to the moldic nature of the molluscs at Belgrade. Therefore, it was deemed practical to establish two names of formational rank for these geographically isolated and lithologically dissimilar units, although both formations correlate in time.

Distribution. The Belgrade and Silverdale formations appear to lie within a structural basin delineated by the Neuse fault to the northeast and the New River. They do not appear to crop out naturally, and exposures are apparently restricted to active quarries.

Correlation. Traditionally, exposures at the Martin Marietta quarry at Belgrade (ON-5) and several small quarries near the town of Silverdale (ON-6) have been referred to the Trent Formation (Kellum, 1925; 1926; Richards, 1943; 1948; 1950; Smith, 1959) (Figure 2).

Several authors have described fossils from the Belgrade Formation exposed at the Martin Marietta quarry at Belgrade (ON-5) (Richards, 1948; Smith, 1959; Lawrence, 1976). In the lower quartz arenite facies, abundant Anomia and barnacles occur. The barnacles belong to the Balanus concavus stock which represent an advanced group characterized by the presence of both inner and outer laminae. The Solidobalanus (Hesperibalanus) barnacles lack the inner laminae. Thus, the Belgrade Formation barnacles do not correlate with the barnacles from the type section of the Trent Formation. The Balanus concavus group is not known from rocks older than Miocene in North America. As indicated in Lawrence (1976), the Belgrade beds are considered late Oligocene by many paleontologists; however, the apparent extensive

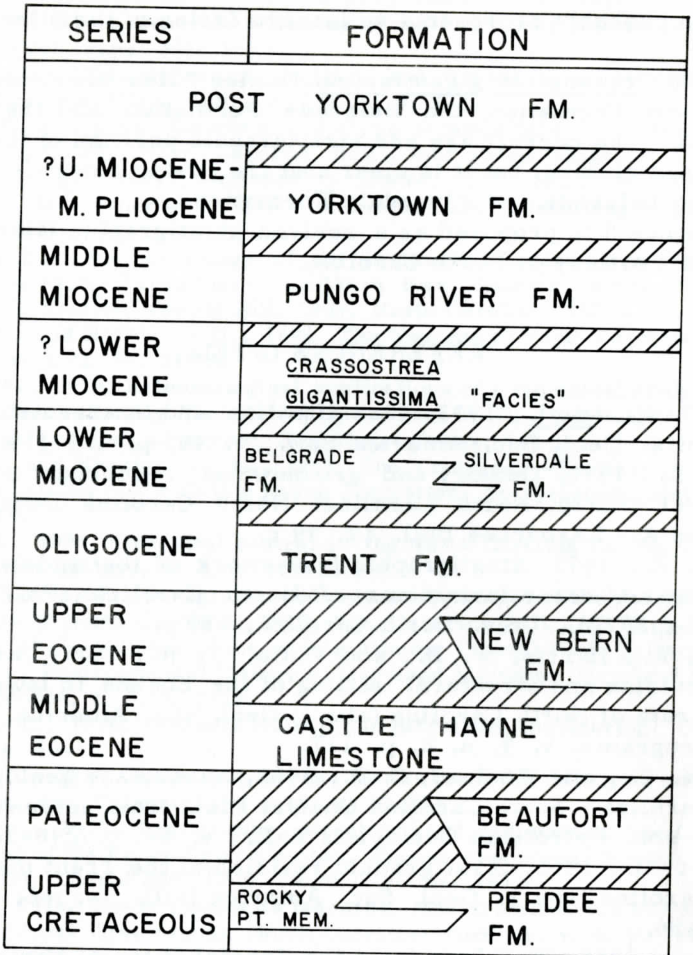

Figure 5. Proposed stratigraphic revisions of the Tertiary of North Carolina.

eustatic sea level drop in the mid-Oligocene (Vail, 1976; Gussow, 1976) would seem to preclude upper Oligocene sediments in the Coastal Plain. Cores drilled at the base of the Belgrade quarry reveal little owing to poor recovery in unconsolidated sediments; however, they indicate a phosphatized surface several meters below the quarry floor. Also, the samples indicate that the Belgrade beds lie disconformably on the barnacle, pelecypod-mold biosparrudite facies of the Trent Formation.

Downdip from the Belgrade Formation, the Silverdale Formation is exposed. The early Miocene age of the Silverdale Formation is well documented (Kellum, 1925; 1926; Richards, 1948; Vokes, 1967; 1970). The barnacles that occur in the Silverdale Formation are conspecific with the barnacles found in the Belgrade Formation. Thus, the Belgrade

Formation probably represents an inshore facies of the Silverdale Formation.

The Crassostrea gigantissima "facies" lies disconformably on the New Bern Formation, the Belgrade Formation and the Silverdale Formation. The correct age and stratigraphic position of these oyster channels are elusive, but it is clear that the oyster channels are younger than the Belgrade and Silverdale Formations.

Figure 5 is proposed as a revised stratigraphic framework for part of the Tertiary of North Carolina.

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PETROLOGY AND BULK ROCK GEOCHEMISTRY OF THE FRANK ULTRAMAFIC BODY, AVERY COUNTY, N. C. AND ASSOCIATED OTHER ULTRAMAFIC ROCK BODIES OF THE SOUTHERN APPALACHIANS

By

John R. Carpenter
and
Hsien S. Chen
Department of Geology
University of South Carolina
Columbia, South Carolina 29208

ABSTRACT

Petrologic and bulk chemical analyses of several samples of the partially serpentinized Frank, N. C. dunite demonstrate generally remarkably similar mineral and chemical compositions. Primary minerals include olivine (generally polygonal and strain free), chromite (as disseminated euhedral to subhedral, and occasional grains of enstatite (generally highly strained). Secondary alteration or hydration products include serpentine, talc and chlorite. Analysis of one sample of orthopyroxenite demonstrated highly strained orthopyroxene and its alteration product talc.

Similar analyses of single samples from several other partially to fully serpentinized ultramafites yielded generally olivine-rich (and orthopyroxene-poor) rocks, similar in mineral and chemical composition to the Frank samples.

These results support the hypothesis that the olivine-rich ultramafites were: (1) emplaced as differentiated ultramafite, (2) probably similar in composition prior to tectonic recrystallization; (3) recrystallized to their present polygonal form.

INTRODUCTION

Bulk rock analyses were performed on eight samples collected from the Frank, N. C. ultramafic body and one single sample collected from six other partially serpentinized dunite bodies to determine if any significant petrologic and/or bulk rock compositional variations exist either <u>within</u> a given ultramafic body or <u>between</u> ultramafic bodies.

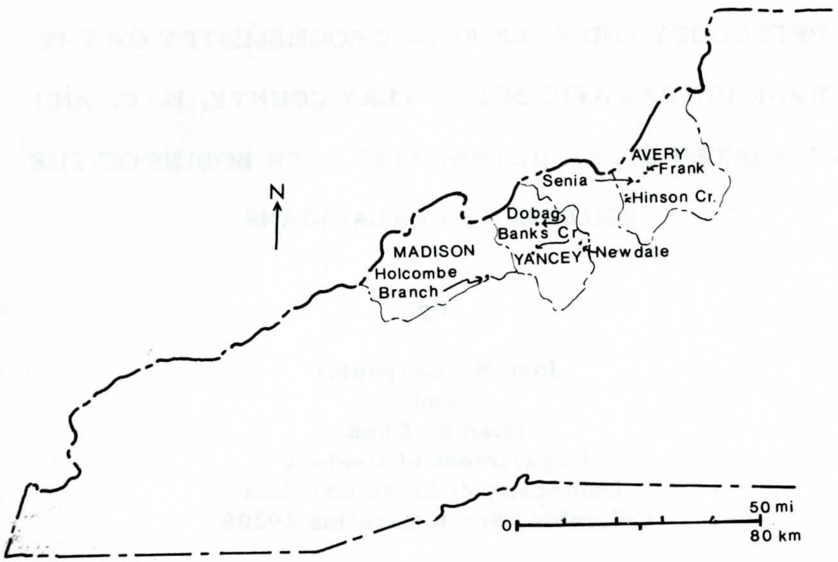

Figure 1. Location map.

For comparative purposes, one sample collected from a completely serpentinized dunite was also analyzed. Samples were taken from the Frank, Senia and Hinson Creek bodies of Avery County, N. C., the Dobag and Newdale bodies of Yancey County, N. C. and the Holcombe Branch body of Madison County, N. C. (Figure 1).

The samples were analyzed for all major constituents and important minor constituents by x-ray fluorescence. Analyzed U. S. G. S. standards AGV-1, BCR-1, DTS-1, G-2, GSP-1, and PCC-1 were the bases for establishing working curves. To obtain points higher in MgO than DTS-1 artificial standards composed of DTS-1 and pure MgO were also analyzed. Ferrous iron oxides were determined by colorimetric titration. H_2O^- and H_2O^+ were determined by weight loss at 105°C and by ignition loss at 1000°C respectively.

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RESULTS

Petrographic examination of the serpentinized dunite shows there to be generally greater than 90 percent olivine, usually as equant, polygonal, and strain-free grains whose size varies of about 0. 1 mm to 1. 0 mm. Other minor, accessory or secondary minerals include

Table 1. Bulk Rock Analyses.

	H.B.	HC-4	HC-5	HC-6	N.D.	S	Mean Burro Mt. Harzburgite	Mean Burro Mt. Dunite
SiO_2	38.53	41.50	41.72	42.24	39.90	39.48	42.97	38.22
TiO_2	.03	.05	.07	.06	.01	.03	.01	.01
Al_2O_3	.20	.34	.42	.43	.31	.37	.93	.61
Fe_2O_3	2.32	1.33	2.31	1.31	2.89	3.76	1.32	2.23
FeO	5.58	5.90	5.90	6.14	4.81	8.11	6.43	5.18
MgO	46.40	49.26	46.96	46.70	47.89	44.46	44.95	47.03
CaO	.19	.06	.08	1.44	.02	.19	.59	.00
MnO	.12	.13	.12	.11	.13	.15	.12	.11
NiO	.31	.36	.34	.35	.35	.31	.31	.35
K_2O	.02	-	-	.01	.04	.04	-	
H_2O^+	6.03	.87	1.93	1.01	2.70	2.91	1.89	5.42
H_2O^-	.20	.11	.12	.03	.20	.16	.10	.22
Total	99.93	99.91	99.97	99.83	99.25	99.93		
Fe_2O_3T	8.52	7.89	8.87	8.13	8.24	12.77		

	B.C.	D	F-1	F-2	F-3	F-4	F-5	F-6	F-7	Mean Frank
SiO_2	40.15	41.86	41.18	42.04	42.64	42.04	43.16	41.60	40.44	41.87
TiO_2	.19	.08	.03	.02	.01	.07	.04	.04	.02	.03
Al_2O_3	.56	1.07	.97	1.05	.56	1.14	.73	.88	1.22	.80
Fe_2O_3	7.20	2.73	2.95	3.61	3.13	2.68	2.50	2.81	5.00	3.24
FeO	1.57	5.45	4.87	4.02	4.88	5.65	5.13	5.03	3.36	4.71
MgO	38.12	43.60	44.60	41.38	44.25	43.65	43.80	44.26	39.31	43.04
CaO	.12	.47	.09	.09	.09	.52	.63	.50	.10	.29
MnO	.33	.31	.31	.29	.32	.30	.28	.30	.28	.30
K_2O	.04	.06	.02	.01	-	.02	.02	.01	.03	.02
H_2O^+	11.57	4.00	4.51	7.07	3.78	3.82	3.40	4.71	9.83	4.29
H_2O	.13	.18	.21	.49	.19	.31	.24	.39	.62	.35
Total	100.06	99.94	99.85	100.19	99.97	100.34	100.05	100.66	100.25	99.08
Fe_2O_3T	8.95	8.79	8.36	8.08	8.55	8.96	8.20	8.40	8.73	8.47

DTS-1	Recom. Value	This Work
SiO_2	40.66	40.72
TiO_2	.01	.03
Al_2O_3	.29	.29
Fe_2O_3T	8.59	8.58
MgO	49.75	49.75
CaO	.14	.15
MnO	.13	.13
NiO	.31	.31
K_2O	.00	.01

HB - Holcombe Branch, N.C.

HC - Hinson Creek, N.C.

ND - Newdale, N.C.

S - Senia, N.C.

BC - Banks Creek serpentinite

F - Frank, N.C.

D - Dobag, N.C.

chromite, which usually occurs as euhedral to subhedral grains approximately 0. 1 mm to 1. 0 mm, talc as isolated grains apparently replacing enstatite, serpentine as alteration products on the edges of and in cracks within olivine, and occasionally a grain or two of enstatite and clinopyroxene.

The chemical analyses of the partially serpentinized dunite samples from the various bodies are given in Table 1. With one exception (to be discussed later) all samples from the Frank body are remarkably similar in bulk rock composition, in terms of the major elements Si, Al, Fe and Mg, and the minor elements Mn, Ni, and Ti. CaO values vary significantly from sample to sample and probably represent a variable, but consistently small, amount of clinopyroxene. Further, the average composition of the Frank samples is remarkably similar (except for Ca) to the compositions of samples from the other dunite bodies. We have also listed the average compositions for dunite and harzburgite from the Burro Mt., California, peridotite and its average composition also is remarkably similar to the compositions of the Southern Appalachian bodies.

Petrographic and x-ray examination of one Frank sample (F-8 in Table 2) showed it to be an orthopyroxenite, composed principally of enstatite, with minor anthophyllite and chromite. The amount of orthopyroxenite in the Frank body, and its relationship to the olivine-rich rocks was not established during the sampling and unfortunately cannot now be determined due to recent extensive quarrying. Orthopyroxenite in the Burro Mt. peridotite, also a alpine-type occurs as thin sills and dikes; in all likelihood, the orthopyroxenite in the Frank body also occurs as thin sills or dikes. The individual enstatite grains are anhedral, have a highly variable grain size, and are uniformly highly strained. The bulk rock composition of F-8, along with the composition of an orthopyroxenite from the Burro Mountain, California (Loney and others, 1971) are all given in Table 2. As can be seen from Table 2 the compositions of the two orthopyroxenites are quite similar with respect to their major and most minor elements. The variation in CaO is probably due either to variable Ca in the orthopyroxene, calcic clinopyroxene exsolution lamellae, or to a small amount of clinopyroxene.

DISCUSSION

The data presented in this paper, taken by themselves, neither support nor refute any of the existing hypotheses concerning the origin and emplacement of these alpine-type ultramafic rock bodies. However, the presence of pyroxenite in the Frank body places some constraints on possible hypotheses. A slightly different picture of the origin and emplacement than the one given earlier by Carpenter and Phyfer (1969) is taking form and this picture is one in which magmatic differentiation (probably within the upper mantle) preceded emplacement into the crust. During emplacement of the ultramafic rocks, the olivine-rich rocks moved either as a crystal mush or as serpentine. The enstatite-rich rocks apparently were emplaced as a solid which underwent considerable deformation. The remarkable similarity in the bulk rock compositions of the partially serpentinized dunites within the Frank body and among the different ultramafic rock bodies basically reflects the also remarkably similar olivine compositions within and among the various ultramafic bodies (Carpenter and Phyfer, 1975). This pattern of consistent bulk rock compositions argues strongly for similar compositions prior to recrystallization.

Considering now the entire Frank body as an entity, it is

Table 2. Orthopyroxenite Analyses.

	F-8	Burro Mt.
SiO_2	55.20	54.65
Al_2O_3	0.08	1.37
Fe_2O_3	1.90	0.53
FeO	4.81	5.03
MgO	32.70	35.02
CaO	0.42	1.19
MnO	0.15	0.13
NiO	0.15	0.12
TiO_2	0.09	0.03
Fe_2O_3T	7.25	6.11

F-8 - Orthopyroxenite from the Frank Body.
Burro Mt. - Orthopyroxenite from Burro Mt., California (Loney and others, 1971).
Fe_2O_3T Total iron as Fe_2O_3

postulated that the ultramafic rock was emplaced either: (1) as a differentiated dunite and pyroxenite body or (2) as a differentiated serpentinized dunite and non-hydrated, unaltered pyroxenite. In either case, either during or after emplacement the olivine recrystallized (from earlier olivine or from serpentine) and the enstatite, for reasons which are still unclear, did not recrystallize. Thus the olivine is interpreted as a product of crustal recrystallization whereas the enstatite probably represents a relict, highly strained, phase of the upper mantle. This interpretation is corroborated by the presence of highly strained isolated enstatite grains in several of the other dunites, presumed also to be relict phases.

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ACROTHORACIC BARNACLE BORINGS FROM THE CHESTERIAN OF EASTERN KENTUCKY AND ALABAMA

By

Frank R. Ettensohn
Department of Geology
University of Kentucky
Lexington, Kentucky 40506

ABSTRACT

Four new Chesterian occurrences of acrothoracic barnacle borings are reported from the Glen Dean Member of the Newman Limestone in eastern Kentucky and from the Bangor Limestone in northwestern Alabama. The easily identifiable, elongate-elliptical borings are only a few millimeters long and have been found exclusively within fragmental bryozoan, crinoid, and brachiopod skeletal material at the four localities. Although the shell-material of the host was apparently abandoned and dead before boring, there was seemingly a high specificity for boring thick bryozoan zoaria at the Glen Dean localities and thick crinoid plates at the Bangor locality.

INTRODUCTION

Although many modern organisms are known to bore calcareous substrates (see Menzies, 1957; Warme, 1975), the identification of these organisms from their borings in the fossil record is generally difficult. The August, 1969 issue (v. 9) of American Zoologist is devoted entirely to such organisms and their borings in calcareous substrates. One group of organisms which produces easily identifiable borings are the acrothoracic barnacles. Although commonly mistaken for annelid worm borings (e. g. Newell, 1942; McKinney, 1968), the small elongate-elliptical borings in calcareous shell material (Figure 1) are known in rocks and sediments from the Devonian through the Recent (Rodriguez and Gutschick, 1970; Tomlinson, 1969a). This report describes two new Chesterian occurrences of barnacle borings from the Glen Dean Member of the Newman Limestone in east-central Kentucky and from the Bangor Limestones of northwestern Alabama.

Acknowledgments

I thank Harrell Strimple for bringing the Alabama specimens to my attention, and I thank Thomas G. Roberts for his critical review of

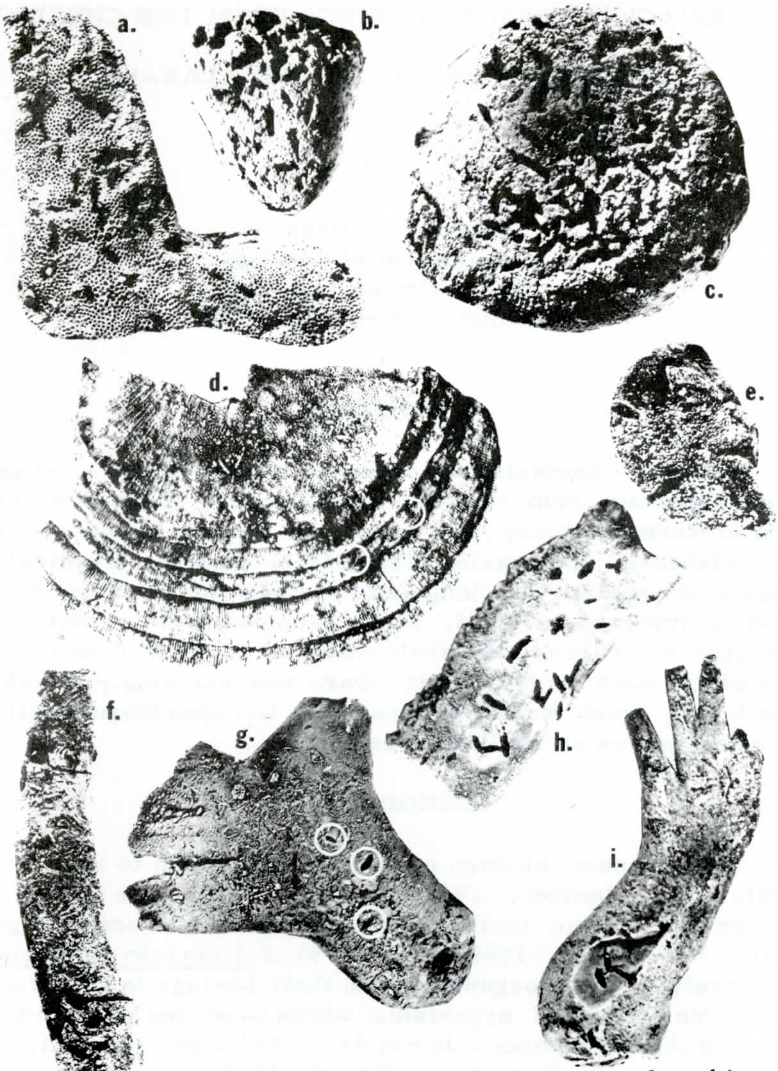

Figure 1. Acrothoracic barnacle borings in bryozoan, brachiopod, and crinoidal skeletal material from the Glen Dean Member of the Newman Limestone and the Bangor Limestone: a, borings in ramose bryozoan, Glen Dean, X 2. 3 (UK 15555); b, borings in infrabasal cone of Agassizocrinus conicus, Bangor, X 2. 9 (SUI 44106); c, borings in basal plate of Staphylocrinus bulgeri, Bangor, X 2. 4 (SUI 44107); d, borings (circled) in pedicle valve of Orthotetes kaskaskiensis, Glen Dean, X 1. 5 (UK 15556); e, four borings seen in transverse section of calcified lateral margin of the bryozoan Lyropora sp. , Glen Dean, X 4. 2 (UK 15557); f, borings in calcified lateral margin of Lyropora sp. , Glen Dean, X 2. 6 (UK 15558); g, borings (circled) in wing plate of

the manuscript. Les Booth did the photography and drafting in Figure 1 and 2. The figured specimens are reposited in the paleontologica. collections of the University of Kentucky (UK) and the University of Iowa (SUI).

OCCURRENCES

Barnacle borings from the Glen Dean Member are known from three localities in east-central Kentucky: 1.) an outcrop on Kentucky State Highway 1274 in Rowan County (Carter coordinates, 1250' FNL x 2600' FSL, 12-S-73), 2.) the Laurel Stone Company quarry in Laurel County (Carter coordinates, 1850' FWL x 300' FNL, 3-H-63), and 3.) the Strunk Construction Company quarry in Tateville, Pulaski County (Carter coordinates, 1900' FWL x 1800' FSL, 15-F-60). Barnacle borings from the Highway 1274 cut were found only with platy, bifoliate bryozoan zoaria, whereas those from the Laurel quarry were found within ramose bryozoan zoaria (Figure 1a), within the heavily calcified lateral margins of the lyre-shaped fenestrate bryozoan *Lyropora*, and within the disarticulated wing plates of *Pterotocrinus* sp. (Figure 1i), *P. acutus*, and *P. bifurcatus* (Figure 1g). Barnacle borings from the Strunk quarry were found within the lateral margins of *Lyropora* zoaria (Figure 1e, f, h), within the infrabasal cones of *Agassizocrinus conicus*, and within one pedicle valve of the brachiopod *Orthotetes kaskaskiensis* (Figure 1d).

Specimens from the Bangor Limestone were found by Harrell and Christina Strimple about two miles southeast of Littlefield, Colbert County, Alabama (SW 1/4, SW 1/4, Sec. 36, T. 5 S, R. 11 W). Barnacle borings from this locality were found within the disarticulated basal, and infrabasal plates of the crinoids *Agassizocrinus conicus* and *Staphylocrinus bulgeri* (Figure 1b, c).

DISCUSSION

The barnacle borings described in this report were found primarily within bryozoan zoaria and crinoid calyceal plates. The borings within the figured specimens (Figure 1) range from 0.7 to 2.2 mm long and from 0.3 to 0.9 mm wide. The borings are largely concentrated on

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Figure 1 (continued)

Pterotocrinus bifurcatus, Glen Dean, X 2.5 (UK 15559); h, borings in calcified lateral margin of *Lyropora* sp., Glen Dean, X 2.7 (UK 15557); i, borings in wing plate of *Pterotocrinus* sp., Glen Dean, X 2.1 (UK 15560).

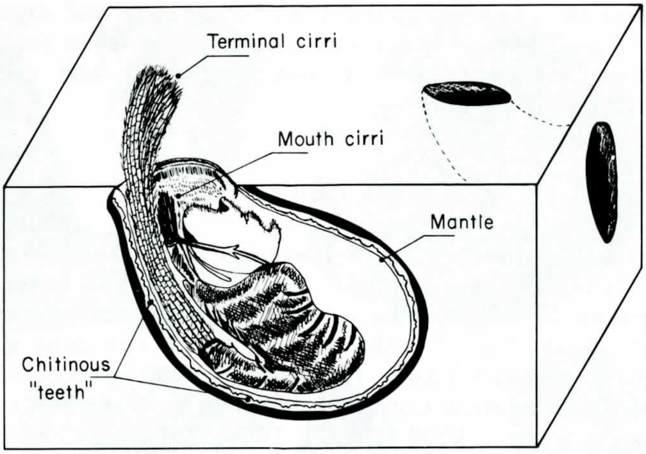

Figure 2. Sectional drawing of an acrothoracic barnacle within its boring into a calcareous substrate (after Seilacher, 1968, and Tomlinson, 1969a).

one surface in each specimen, though randomly oriented on that surface. The distribution of borings suggests that the barnacles penetrated the upper exposed surface of dead, disarticulated skeletal debris lying on the substrate. Abandoned shells of dead animals were apparently favored by the barnacles in most examples of fossil and recent borings, because of the absence of periostracum or other organic tissue covering the shell (Rodda and Fisher, 1962; Tomlinson, 1969a, b). Seilacher (1968, 1969) however, contends that some barnacle borings penetrated the shells of live hosts in such a way as to take advantage of currents generated by the feeding or locomotary activities of the host.

Fossil barnacle-borings are known from the shell material of corals, brachiopods, bryozoans, various echinoderms and various molluscs. Despite the variety of invertebrate host material available for boring, Rodda and Fisher (1962) noted a high host specificity at a given locality. A similar host specificity is apparent in the Glen Dean and Bangor. Most of the borings from the Glen Dean occur in thick bryozoan zoaria (Figure 1a, e, f, h), although scattered borings are found in crinoid plates and brachiopod valves (Figure 1d, g, i). In the Bangor Limestone, the thick calyceal plates of stemless crinoids were apparently preferred (Figure 1b, c). Such host specificity is progably related to a preference for boring the thickest skeletal material available (Tomlinson, 1969a).

Because these small barnacles lack any protective exoskeleton, boring into calcareous substrates provides a means of protection; little if any harm is done to a live host when present (Tomlinson, 1969a, b). Tomlinson (1969a) indicates that boring is accomplished through abrasion by chitinous teeth extending outward from the mantle surface (Figure 2), although initial larval boring is by dissolution. The mantle is cemented only to the dorsal side of the burrow, which provides a fulcral point for movement of the mantle.

The only part of the barnacle extended outside the burrow are the terminal cirri (Figure 2) which are thrust outward in sweeping arcs. The sweeping action of the terminal cirri sets up currents which carry particulate material into the mantle cavity where the mouth cirri (Figure 2) filter out any food particles. For a more detailed account of the biology of the boring acrothoracic barnacles see Tomlinson (1969a).

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PETROLOGY OF AN ULTRAMAFIC BODY NEAR MICAVILLE,

YANCEY COUNTY, NORTH CAROLINA[1]

By

R. H. Kingsbury
and
R. A. Heimlich
Department of Geology
Kent State University
Kent, Ohio 44242

ABSTRACT

The ultramafic body is ellipsoidal in map view, measures 500 by 950 feet, and is void of mineralogic layering, jointing, and megascopic preferred orientation of grains. The country rock is quartzofeldspathic gneiss characterized by foliation that strikes NE-SW and dips steeply SE. The long axis of the pluton is oriented NNE-SSW; the body is slightly discordant.

Primary minerals of the dunite are magnesian olivine, magnesian enstatite, diopside, and chromite. Olivine is found as equant, polygonal, recrystallized grains possessing triple-point junctions that meet at 120 degrees, and as smaller anhedral strained grains or larger porphyroclasts which exhibit elongation and undulose extinction. Secondary hydrous alteration minerals, serpentine, chlorite, talc, vermiculite, and anthophyllite occur sporadically along the periphery of the body and decrease in abundance toward its interior. Locally silicification of the dunite is significant. Textural and cross-cutting relationships define two distinct, and possibly three stages of hydrous alteration.

Two oriented samples characterized by strained anhedral and elongate olivine grains exhibit a strong preferred orientation with Y and Z girdles normal to an X point maximum. A third sample composed of polygonal olivine grains exhibits a more diffuse oriented fabric.

Analysis of the data indicates that the pluton was emplaced as a cold solid mass of deformed mantle-derived ultramafic material which underwent subsequent hydrous alteration. Development of the secondary minerals was accompanied by removal of magnesia from the dunite and introduction of silica, alumina, and water presumably from the enclosing country rock.

[1] Contribution No. 148, Department of Geology, Kent State University.

INTRODUCTION

Alpine ultramafic bodies distributed along the axis of the Appalachian Mountains are currently being studied because of their significance with respect to continental drift and plate tectonics. Based on lithology and field relationships two distinct mechanisms of emplacement have been postulated for these bodies associated with folded mountain belts. In Newfoundland, several ultramafic masses have been studied extensively and identified as fragments of oceanic lithosphere obducted laterally onto continental margins (Church and Stevens, 1971; Coleman, 1971; Chidester and Cady, 1971; Dewey and Bird, 1971). Other bodies, both in Newfoundland and in the southern Appalachians appear to have been emplaced vertically above subduction zones as cold, solid masses (Chidester and Cady, 1971; Stevens, Strong and Kean, 1974).

Recently, there has been renewed debate regarding the extensive serpentinization of the ultramafites. Citing petrofabric and geochemical data, some workers argue for emplacement as serpentine which later dehydrates to olivine (Carpenter and Phyfer, 1976; Hearn et al., 1977). Others believe the bodies were emplaced as fresh ultramafic material (Sailor and Kuntz, 1973; Greenberg, 1976; Bluhm and Zimmerman, 1977) which later underwent serpentinization (Hahn and Heimlich, 1978; Dribus et al., 1977).

Because of the current interest in plate tectonics and mantle processes, it has become obvious that there is a need to re-examine these alpine ultramafic bodies to determine the source of the ultramafic material, their mode of emplacement, their recrystallization and hydrous alteration histories, and their role in the evolution of the continents.

The ultramafite under discussion is one of a number of such bodies present in a belt that extends NE-SW in western North Carolina, within the Blue Ridge Province of the Appalachian Mountains. It is located in Yancey County, approximately 4.5 miles north of Micaville at the site of the J. C. Woody Asbestos Mine (Figure 1). Other than a brief description (Conrad et al., 1963) of associated anthophyllite asbestos, no literature was found on this body.

Acknowledgments

We are pleased to acknowledge the Society of the Sigma Xi which provided financial assistance for this study. Special thanks go to J. C. Woody for help, hospitality, and permission to work on his property, and also to Tim Boyd for valuable assistance during the course of the field work. As well, we thank E. H. Carlson, and D. F. Palmer for their advice during the laboratory studies, and for critical reading of an earlier draft of the paper.

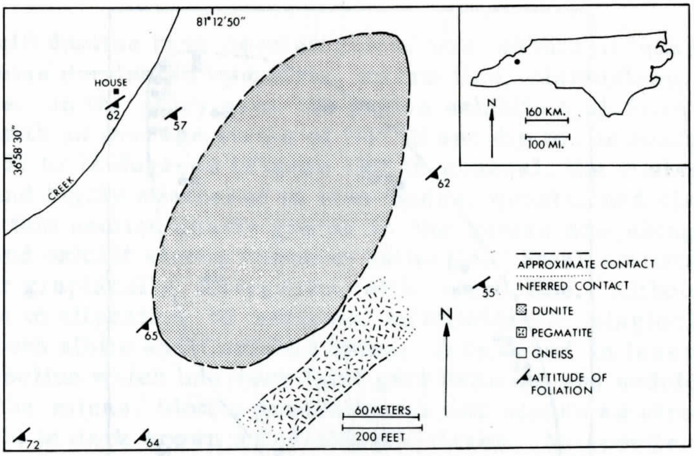

Figure 1. Geologic map of the dunite.

METHODS OF INVESTIGATION

Sampling of the dunite was restricted to the periphery due to lack of outcrop in the interior. All areas of outcrop were sampled and, where previous mining operations provided substantial exposure, samples were collected on roughly 20-foot intervals and located with the aid of a steel tape (Figure 2). Because of generally poor exposure, relatively few samples of the country rock were obtained. All sample localities were plotted on a base map (scale 1:100) constructed by pace and compass technique.

Thin sections were prepared from selected samples and examined to determine mineral content, mineral percentages, and textures using a Leitz petrographic microscope. Modes of 1200 points were obtained for each thin section. Petrofabric analysis was conducted on three oriented samples using a five-axis universal stage following the procedure outlined by Emmons (1943). Data from 150 grains per thin section was plotted on an equal area stereographic net and values were contoured at intervals of 1, 3, 5, 7, 9, and 11 percent.

Forsterite content of eight olivine separates was determined, with $\pm$ 3 percent error, by x-ray diffractometry following the method described by Yoder and Sahama (1957). Silicon was used for an internal standard in the determination except where interferences from alteration minerals required the use of lithium fluoride.

Pyroxene, plagioclase, and alteration products were identified by critical optical properties. Plagioclase composition was determined by measurement of extinction angles of albite twins using the Michel-Levy determinative chart (Kerr, 1959). Pyroxene composition was estimated by 2V measurements on the universal stage and use of the determinative chart given by Poldervaart (1950).

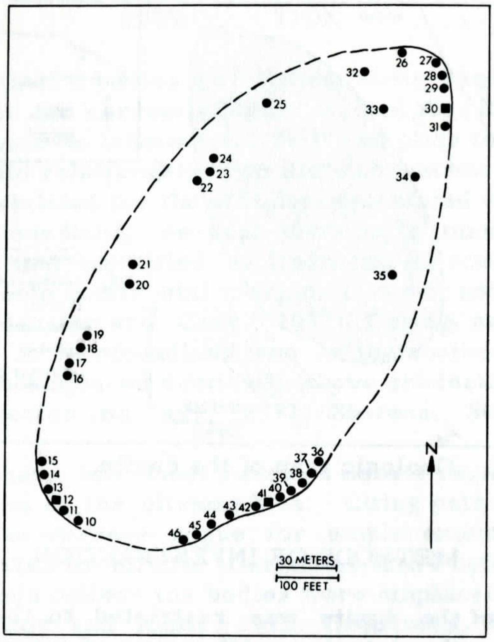

Figure 2. Sample locality map. Squares are oriented-sample localities.

Chemical analysis of selected samples was performed for major and several trace elements using the Perkin-Elmer 403 atomic absorption spectrophotometer. Accuracy was determined by comparing values obtained from U. S. G. S. whole rock standards DTS-1 and PCC-1 with known values. Duplicate analyses were run for each sample to check precision which was better than 5 percent.

GEOLOGIC SETTING

Covered with dense vegetation, the dunite forms a resistant topographic high. In map view, the body is ellipsoidal and roughly 500 by 950 feet in maximum dimensions. Although its long axis is nearly coincident with the strike of foliation in the country rock, its contact, at the north and south ends, transects the foliation (Figure 1). There is no evidence of deflection of country rock structure by the body. The dunite lacks jointing, mineralogic layering, and megascopic preferred orientation of grains. Along the contact with the country rock the dunite is altered extensively.

Another small ultramafic body occurs approximately 450 feet southwest of the main body. It measures 20 by 45 feet and is transected by a vertical six-inch thick vein of anthophyllite.

Both dunites have been emplaced in a terrain of quartzo-feldspathic gneiss dominated by quartz, microcline, plagioclase, muscovite and biotite. In the study area the gneiss exhibits a uniform attitude of foliation with an average strike of N49^{o}E and dip to the southeast ranging from 57 to 72 degrees (Figure 1). In general, the gneiss is poorly exposed and highly weathered to iron oxides, quartz, and clay.

In thin section quartz grains in the gneiss are elongate, xenoblastic, and exhibit strong undulose extinction. Small amounts of quartz also occur graphically intergrown with microcline. Although slightly cloudy due to alteration to sericite, subidioblastic plagioclase (An32) displays both albite and Carlsbad twins. It is found in lesser amounts than microcline which has very faint grid twinning and undulose extinction. Of the micas, biotite predominates and occurs as strongly pleochroic light to dark brown subidioblastic flakes. Muscovite, also subidioblastic, is most commonly found associated with biotite. Accessory minerals include epidote, zircon, apatite, and magnetite. Zircon occurs as xenoblastic grains commonly as inclusions in quartz and biotite. Epidote, which possesses well developed cleavage, and apatite are subidioblastic. Magnetite is most commonly found along grain boundaries as small subidioblastic grains. The sub-parallel alignment of elongate and micaceous minerals defines a lepidoblastic texture in the gneiss.

A granitic pegmatite, composed of quartz, orthoclase, and muscovite, is located on the southeast side of the pluton (Figure 1). It is approximately 100 feet thick, crops out over a distance of 400 feet, and appears to be concordant with respect to foliation in the enclosing gneiss. The orthoclase is white and highly weathered to kaolinite for which the pegmatite was mined (Conrad et al, 1963).

DUNITE BODY

Petrography

The pluton is composed of enstatite-bearing dunite in which olivine, enstatite (0-22.3%), diopside (0-10.7%), and chromite (0.5-7.2%) occur as the primary minerals (Table 1). Olivine occurs typically as equant polygonal grains or as irregularly shaped grains. The irregular grains range in diameter from 0.2 to 3.0 mm. and show varying degrees of alteration with some grains almost totally serpentinized. Most exhibit deformational features such as undulose extinction, faint strain bands, and granulation. Strained porphyroclasts as large as 0.4 by 4.0 mm. are typically elongate, fractured, and surrounded by mosaics of small equidimensional recrystallized grains ranging in diameter from 0.1 to 1.0 mm. The recrystallized grains lack deformational features, have polygonal shapes, and possess triple-point junctions which meet at 120 degrees, textures common to recrystallized rocks (Ragan, 1969; Alcorn and Carpenter, 1976). Polygonal grains,

Table 1. Modal and mineralogic data for the dunite.

Sample Number	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26
Olivine		87.4	72.5	63.3	55.6	59.8	53.5	56.6	61.6	60.0	35.8	47.8	45.7	54.2	40.0	19.7	
Enstatite		3.4	18.9	8.3	5.4	0.8	16.0	13.6	4.7	17.3	15.9	7.0	14.2	12.6	22.3	18.9	
Diopside																8.6	
Chromite	1.7	1.2	1.3	1.8	2.2	4.0	0.8	0.5	1.2	1.0	1.3	0.6	1.6	2.6	1.4	1.9	7.2
Serpentine		3.4	3.1	15.8	25.7	22.0	21.0	19.0	27.4	14.5	28.8	35.1	27.6	18.0	25.4	4.7	0.6
Anthophyllite	92.2				0.5											21.6	34.4
Chlorite	2.0	4.0	3.9	10.0	10.1	13.5	8.0	9.2	3.0	5.2	8.8	8.5	8.1	11.4	9.4	7.5	1.8
Talc		0.3	0.2	0.7	0.3		0.6	1.2	2.0	2.0	9.3	0.8	2.0	1.0	1.5	14.1	
Vermiculite	4.0															1.4	55.9
Chert																	
Hematite					tr												
Olivine (Fo%)			89.3				90.4					90.2		88.8			
Enstatite 2V			88°				85°					79°		86°			
Enstatite ($MgSiO_3$%)			87				89					92		88			

Sample Number	27	29	29A	30	31	32	33	33A	35	38	38A	41	42	43	44	45	46
Olivine		19.0		31.1	40.8	38.0	43.0		30.0	47.6		82.0					
Enstatite	7.2	8.3	4.7	16.0	5.4				1.7	2.0		5.8					
Diopside	6.1	1.1	9.6											10.7			
Chromite	1.1	2.3	0.5	1.9	3.8	5.0	2.2	*44.4	0.8	2.5	1.7	1.2	4.0	2.0	21.5	3.3	tr
Serpentine		36.0		31.6	34.8	40.0	43.8		56.0	17.2							
Anthophyllite	83.0	20.9	68.4	10.4	5.7				0.5			3.4		56.4		39.3	19.5
Chlorite	1.0	4.4	16.8	6.5	8.7	16.7	11.8	55.5	9.7	3.8	4.0	6.8	2.6	31.0			
Talc		5.2		1.2	0.4				0.8	0.4	30.6	0.6	18.3				
Vermiculite	1.8	2.5		1.0											78.5	57.4	77.9
Chert										15.7	49.3		58.0				
Hematite				tr						10.6	14.3		17.0				
Olivine (Fo%)						91.6	89.3		91.5			89.8					
Enstatite 2V						87°	83°		83°			85°					
Enstatite ($MgSiO_3$%)						88	90		90			89					

*Magnetite

although commonly bordered by serpentine, are not as thoroughly altered as the large anhedral porphyroclasts. X-ray analysis of eight samples, representative of the periphery of the pluton, indicate that the olivine is magnesium-rich and averages 90% forsterite (Table 1). The range in values for the samples analyzed falls within the 3% limit of error for the method used and therefore no trend in olivine composition is apparent with respect to sample location.

Enstatite occurs as colorless subhedral grains ranging in maximum dimension from 0.2 to 4.0 mm. It possesses well-developed cleavage, and larger grains contain numerous inclusions of rounded olivine grains. Values for 2V range from 79° to 88°. Enstatite is found in varying stages of alteration and some grains exhibit mild undulose extinction. Composition averages 89 weight percent $MgSiO_3$ (Table 1).

Diopside was detected in five of the 34 samples analyzed petrographically. It appears as colorless, non-pleochroic, subhedral grains with poor cleavage and maximum dimensions of 0.2 to 1.0 mm. It is most commonly found in the northern part of the pluton.

Chromite occurs throughout the body as subhedral to subrounded grains 0.1 to 1.5 mm. in maximum dimension. Altered chromite is surrounded by rims of chlorite, and all gradations exist from fresh chromite to patches of chlorite enveloping tiny remnant chromite grains.

Hydrous alteration minerals include serpentine, chlorite, talc, vermiculite and anthophyllite. Serpentinization is extensive in some areas where it exceeds 40% of the rock. Chrysotile dissects and borders grains of olivine and enstatite and forms a mesh-like texture of intersecting veins in extensively altered samples. Antigorite occurs commonly as irregular patches within primary minerals. Tiny subhedral grains of secondary magnetite are found within the serpentine minerals.

Chlorite is an alteration product of all the primary minerals. Small subhedral, non-pleochroic laths are found within olivine and large pyroxene grains. Commonly it is found enveloping and replacing chromite where it occurs as large single laths or as aggregates of smaller subhedral laths.

Highly irregular patches of talc replace both olivine and enstatite. Both talc and chlorite are rare as vein and fracture fillings. In some places they straddle and replace serpentinized grains of olivine and enstatite.

Anthophyllite occurs as relatively unaltered acicular crystals and as fibrous masses. Individual crystals, which range in size from 1 to 6 mm., transect olivine grains and areas of talc and serpentine. With one exception the anthophyllite is restricted to a zone of intense alteration along the northern contact of the pluton.

Locally vermiculite and chert are significant secondary minerals. Vermiculite occurs as pale brown flakes, some of which are bent and show undulose extinction. It is found in large amounts of the

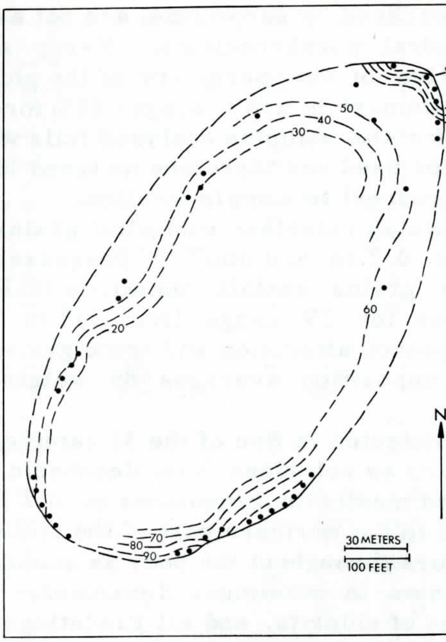

Figure 3. Alteration map.

southeast side of the pluton but is also present along the northern contact where anthophyllite is abundant. Chert is common in parallel veins replacing primary minerals. That possessing faint polygonal outlines is pseudomorphic after recrystallized olivine grains at the southeast contact of the pluton. Hematite occurs almost exclusively in samples that have been partially replaced by chert. It is commonly found filling cracks and surrounding grains in the silicified samples. Elsewhere in the pluton it occurs in trace amounts interstitially and coating primary minerals.

Figure 3 shows the distribution of total alteration minerals across the pluton. Contours were constructed using percentages obtained from thin section modes. Within the area sampled, alteration appears to be most extensive at the contact, decreasing toward the interior. Alteration is particularly intense on the north side of the pluton, but also along the southeast contact where it may have been influenced by the granitic pegmatite.

Chemical Composition

Nine dunite samples were prepared for chemical analysis of the major oxides, silica, alumina, magnesia, and Fe_2O_3, and trace elements nickel and chromium (Table 2). Values were plotted on a map of the pluton and contoured to determine the presence or absence of

Table 2. Chemical analyses of the dunite.

Sample Number	12	16	21	23	26	32	33	35	41
SiO_2	40.95	49.25	49.62	50.03	50.17	41.32	41.71	47.54	41.18
MgO	52.61	43.59	42.34	42.78	35.30	47.23	49.76	45.65	51.84
Fe_2O_3*	5.34	5.35	5.25	5.57	6.51	5.84	4.66	5.36	5.81
Al_2O_3	0.80	0.91	1.55	1.26	5.97	3.85	3.71	0.97	1.08
Cr (ppm)	2460	2765	2785	2692	1254	9365	6161	3329	3175
Ni (ppm)	2352	2297	2294	2343	2414	2277	2406	2223	2322
Totals	99.7	99.1	98.7	99.6	97.9	98.2	99.8	99.5	99.9

*Total Fe as Fe_2O_3

trends for each element as a function of sample location. While chromium, nickel, and Fe_2O_3 do not show detectable trends, silica, alumina, and magnesia exhibit suggestive systematic variations across the pluton. Although data are sparse and samples are lacking from the interior, the magnesia values show a decrease toward the periphery of the pluton whereas silica and alumina values show the reverse. These variations are probably due to reactions between the magnesian dunite and the highly siliceous country rock. They appear to be related to the suite of alteration minerals present at the borders of the pluton. Alteration of the dunite is most intense along the northeast contact where values for magnesia are lowest and those for silica and alumina are greatest.

Petrofabric Data

Petrofabric analysis was conducted on three oriented samples taken from the periphery of the pluton. Two of the samples, located near the northeast (#30) and southwest (#12) contacts, are characterized by strained anhedral olivine grains, some of which show slight elongation. Equant polygonal grains are absent from these two samples. The third sample, near the southeast contact (#41), lacks deformational features and shows evidence of recrystallization due to the abundant polygonal grains with 120-degree triple-point junctions.

Petrofabric data (Figure 4) for the two samples characterized by anhedral olivine grains show X and Z maxima as high as 11% with Y slightly lower at 9%. In these samples (#12, 30) the X axis defines a strong point maximum with Y and Z forming girdles normal to X. The third sample (#41), with polygonal, recrystallized grains, displays a more diffuse pattern for all three axes. The Y-axis shows a weak point maximum and Z forms two point maxima both at levels of 5%. The X-axis generally follows the periphery of the diagram but forms a weak girdle contoured to 3%.

The two distinctly different results from the three samples analyzed could be a function of the recrystallized versus the deformational textures, or the location from which the samples were taken. None of the olivine crystallographic axes in the samples appear to parallel

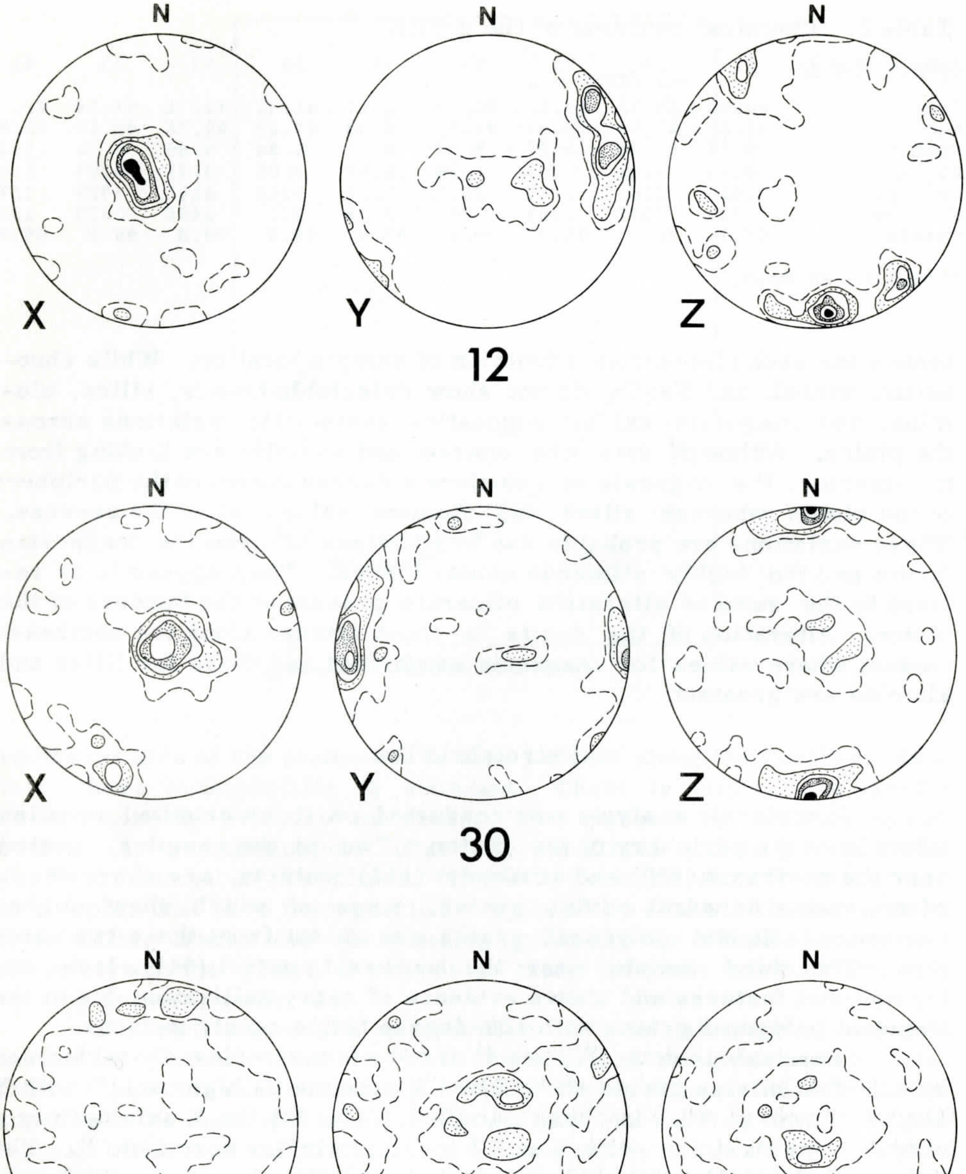

Figure 4. Petrofabric diagrams for oriented samples (Contours at 1, 3, 5, 7, 9 and 11%.

the strike of foliation in the country rock. Therefore, the preferred orientation, clearly characteristic of some of the dunite, appears to be unrelated to regional deformation.

DISCUSSION

The highly magnesian olivine and enstatite, and high nickel and chromium values indicate a mantle source for the dunite (Ragan, 1963). The dunite is also similar in chemical composition, textures, and tectonite fabric to peridotite nodules in basalts, further indication of a mantle origin (Mercier and Nicolas, 1975). Moreover, it resembles other southern Appalachian alpine ultramafic bodies for which a mantle source has been postulated based on field occurrence, textures, fabric, mineralogy, and chemical composition (Dribus et al., 1977; Hahn and Heimlich, 1978; Bluhm and Zimmerman, 1977; Yurkovich, 1976; Palmer et al, 1977).

The dunite textures are deformational in origin. The presence of undulatory extinction and strain bands in many of the olivine grains suggests that the dunite has undergone plastic flowage and recrystallization (Raleigh, 1967 and 1968). The characteristic unstrained equant polygonal grains which meet at 120-degree triple-point junctions and which surround larger deformed porphyroclasts indicate further that the rock has undergone recrystallization (Ragan, 1969). We believe that these recrystallization features were inherited from the mantle source region of the dunite and that they reflect syntectonic recrystallization, the dominant mode of flow in the upper mantle (Ave'Lallemant and Carter, 1970).

In the three samples analyzed, the preferred orientation of the olivine crystallographic axes fails to show a systematic relationship to the attitude of structures in the country rock. Thus it was formed prior to final emplacement of the dunite, presumably in the source region.

The dunite has a complex alteration history that began with an episode of serpentinization with which minor steatization and chloritization were associated. This event produced the serpentine-talc-chlorite-vermiculite alteration assemblage commonly observed. The presence of chlorite flakes and patches of talc cutting and replacing serpentine veins within and along the periphery of primary mineral grains suggests that steatization and chloritization may have been later or of longer duration than the serpentinization. Cross-cutting relationships and the fresh appearance of the anthophyllite suggests it may have formed during a later stage of alteration. Silicification is only locally significant and it may be related entirely to intrusion of the granitic pegmatite. However, the timing of this alteration episode could not be determined because interrelationships with other hydrous alteration minerals were not observed.

The extent of serpentinization varies sporadically across the body, but it is generally greatest at the periphery. This pattern could have developed by hydrous alteration of the ultramafic mass during emplacement, or by dehydration of serpentine beginning at the core of the body and proceeding outward. However, the presence of preferred orientation of grains in the dunite indicates that the body was not emplaced as serpentine and later dehydrated to olivine (Astwood, Carpenter, and Sharp, 1972).

Considerable debate has centered on the problem of constant-volume versus constant-composition during serpentinization (Thayer, 1966, 1967; Page, 1967; Condie and Madison, 1969; Mossman, 1970; Alcorn and Carpenter, 1976). One aspect of the problem stems from the fact that there is a lack of quantitative data concerning the chemistry of serpentinization. For serpentinization to take place at constant-volume, proponents of this view agree that significant amounts of magnesia must be removed from the dunite in solution (Thayer, 1966, 1967; Alcorn and Carpenter, 1976). Unfortunately, field and chemical evidence supporting magnesia migration away from the dunite is generally absent. Mossman (1970) describes radial fracture patterns surrounding partially serpentinized dunite but believes these fractures could develop with very slight volume increases. In support of constant-volume serpentinization, Thayer (1966) examined stratiform ultramafic bodies in which layers of fresh dunite grade into serpentinized dunite with no apparent disruption due to volume changes.

No evidence, such as fracturing or local disruption of the country rock was observed to indicate that the dunite under discussion underwent significant volume increase during serpentinization. However, substantial amounts of magnesia appear to have been removed from the pluton and areas of magnesia depletion generally coincide with areas of extensive serpentinization. Although the debate continues and no widely accepted mechanism for constant-volume serpentinization has been discovered, serpentinization of this dunite appears to have taken place with little or no increase in volume.

Chemical data suggests a relationship between the alteration minerals and the changes in magnesia, alumina, and silica values across the pluton. Systematic variations include increases in silica and alumina and a decrease in magnesia locally from the interior to the periphery of the pluton. On the north side of the pluton where the percentage of alteration minerals approaches 100, silica and alumina values are at a maximum and magnesia values are the lowest. In general, alteration of the pluton occurs with removal of magnesia from the dunite, and introduction of silica, alumina, and water presumably from the enclosing country rock.

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A MARCASITE LAYER IN PRODELTA TURBIDITES OF THE BORDEN FORMATION (MISSISSIPPIAN) IN EASTERN KENTUCKY

By

J. Barry Maynard
Fisk Laboratory of Sedimentology
Department of Geology
University of Cincinnati
Cincinnati, Ohio 45221

and

Sandra K. Lauffenburger
Exxon Production Research Company
Box 2189
Houston, Texas 77001

ABSTRACT

A prodelta turbidite unit in the Borden Formation contains a layer of marcasite in the bottom-most turbidite siltstone. The geometry and extent of individual turbidite beds in this unit have been determined, and it is shown that the mineralization is confined to one bed. It is believed to have developed during early diagenesis at the contact between organic matter-rich pore waters in the mud, and more oxidizing, sulfate-containing pore waters in the permeable siltstone. It is suggested that marcasite forms in preference to pyrite in these rocks when the rate of sulfide precipitation is very slow.

INTRODUCTION

The Borden Formation is a dominantly clastic unit formed by the westward progradation of deltaic sediments during early Mississippian time. It extends from the western part of the Appalachian Basin in eastern Kentucky and Ohio (where it is referred to as the Cuyahoga Group) into the Illinois Basin (Lineback, 1966, 1968). To the east, similar rocks occur in the Price Formation (Kreisa and Bambach, 1973), where the transition from marine to non-marine strata can be seen. Commonly occurring in the Borden and its equivalents are bundles of siltstone beds that are thought to be turbidites (Lineback, 1968; Moore and Clarke, 1970; Weir, 1970; Kepferle, 1977). Siltstones of this type are particularly well-exposed around Morehead in Rowan County, Kentucky. Here the upper Devonian and lower Mississippian consist predominantly of black and gray shales with some siltstone (Figure 1). The turbidite beds occur in the Farmers Member of the Borden Formation and are separated from the underlying black Sunbury Shale by a gray shale unit known as the Henley Bed (Hoge and Chaplin, 1972). The bottom-most turbidite bed in the Farmers is extensively mineralized by marcasite. The purpose of this paper is to describe this mineralization in the context of the local sedimentology of the Borden Formation.

Acknowledgments

This project has greatly benefitted from the help of Paul Potter and Roy Kepferle. Lily Kao made the carbon measurements.

THE BORDEN FORMATION IN THE VICINITY OF MOREHEAD, KENTUCKY

The excellent exposures of the Borden in the study area permit the detailed study of the geometry of individual turbidite beds. These beds were deposited on the eastward-thickening wedge of the Henley gray shale (Figure 2). A typical isopach of an individual bed shows a rounded body covering about 150 km^2, with a ridge extending back in the up-current direction (Figure 3). Thickness ranges from about 10 to 100 cm. In contrast to the siltstone beds, which change thickness rapidly, the intervening shales have a nearly constant thickness, a useful feature for correlation.

Paleocurrents, as revealed by sole markings, are uniformly towards the west (Figure 4) in agreement with Moore and Clark (1970). There are also fewer siltstones in the distal direction.

The Henley Bed, in the study area, can be divided into three subunits: a lower glauconitic greenish shale, a middle dark gray shale, and an upper light gray shale (Table 1). Note that the organic carbon contents (determined using a Perkin-Elmer 240 Elemental Analyzer) parallel the color changes, being low for the upper and lower beds, but relatively high in the middle subunit. Other dark gray shales in the study area (e. g. , Bedford) have organic carbon values in the neighborhood of 0. 2 to 0. 4 percent. By contrast, the black shales range from 5 to 15 percent. To the north of the study area, in the Burtonville Quadrangle, the Henley is buff, and contains two thick zones of red shale. Thus the oxidation state of the original sediments underlying the Farmers turbidites seems to have decreased from north to south, but was never as low as during the formation of the black shales such as the Sunbury and Ohio.

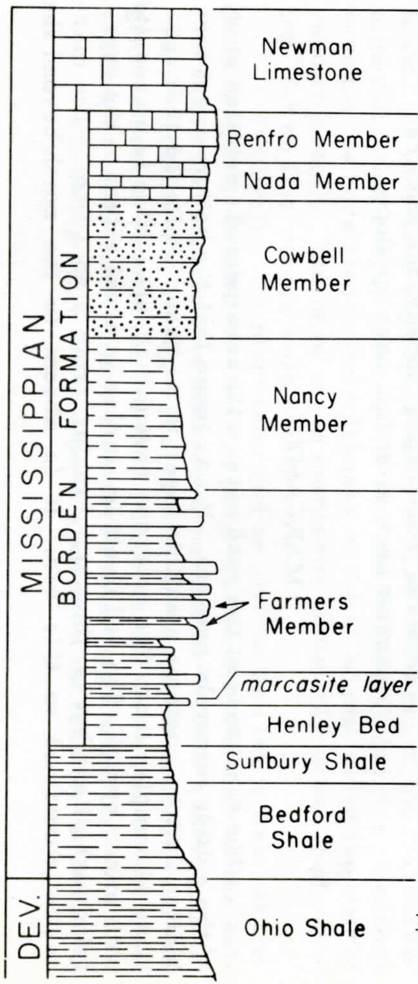

Figure 1. Stratigraphy of the Devonian-Mississippian clastic rocks in the vicinity of Morehead, Kentucky.

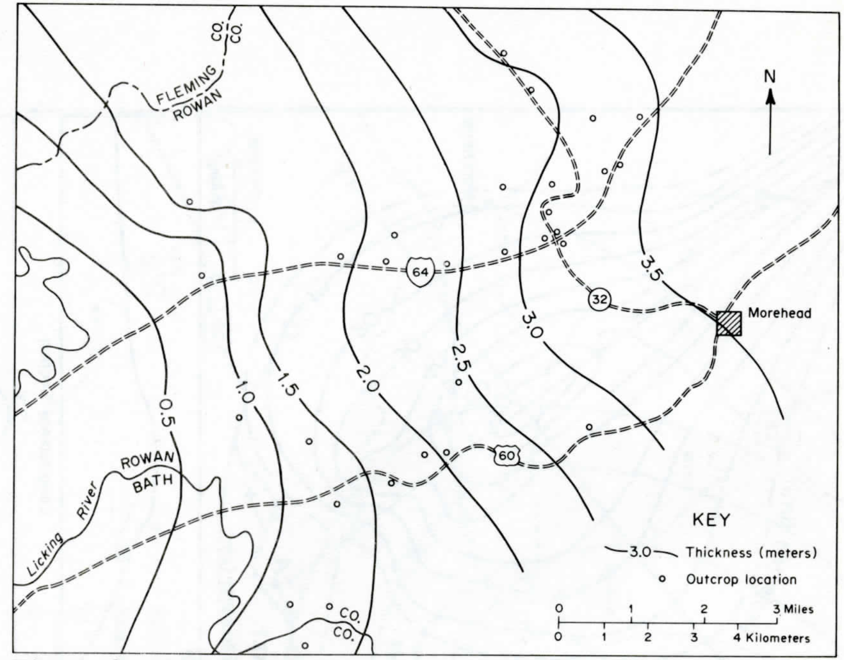

Figure 2. Local isopach map of the Henley Bed. If the top of the underlying Sunbury Shale was essentially flat at the time the Farmers Member was deposited, the isopachs give the slope on which the turbidites were deposited.

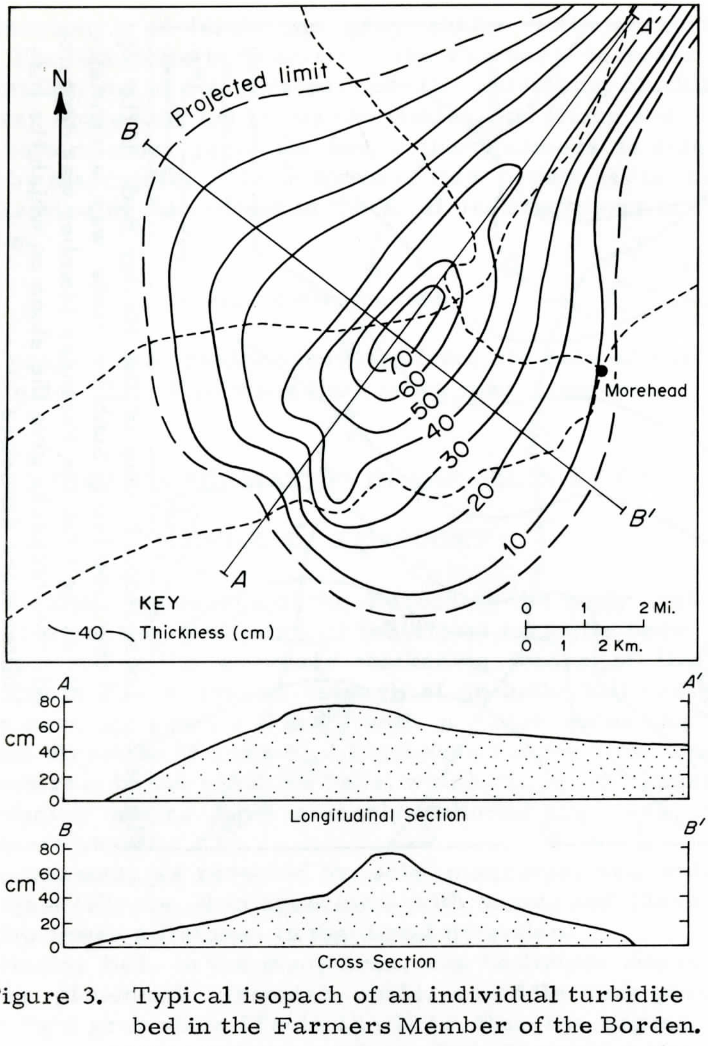

Figure 3. Typical isopach of an individual turbidite bed in the Farmers Member of the Borden. Note the fan-shaped appearance and the narrow neck extending back up slope.

MARCASITE

The freshness of the road cuts in the area permits detailed study of the sulfide minerals present. It was found that the bottom-most turbidite siltstone contains a sulfide layer in its base. Polished thin sections of samples from five outcrops reveal that this sulfide is marcasite with minor pyrite. The bed itself is similar in areal extent to the overlying turbidites, but is thinner, ranging from 2 to 20 cm. The marcasite is confined to this bed: it is absent to the north beyond the

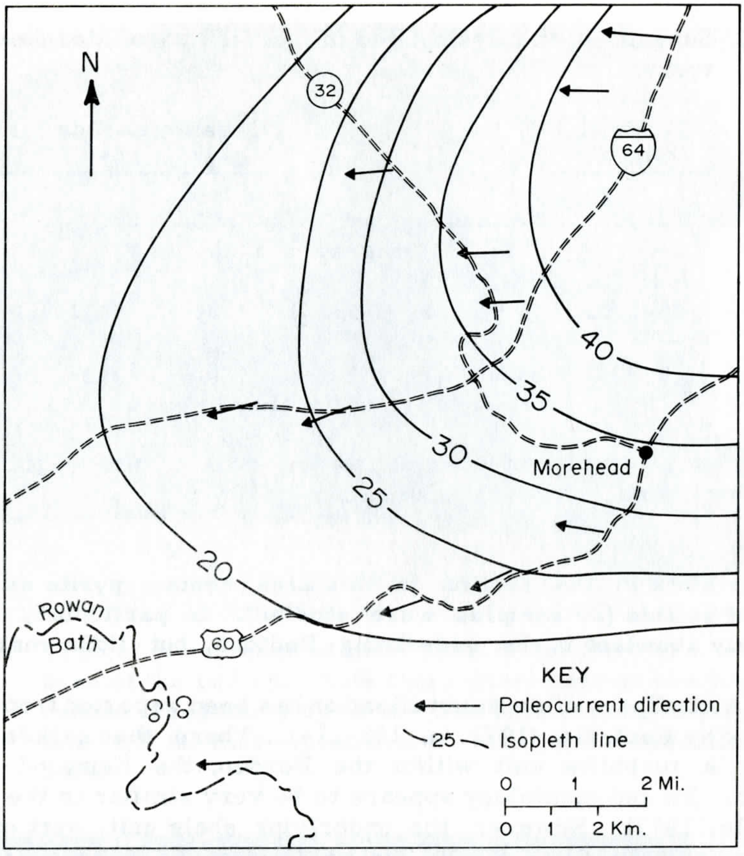

Figure 4. Paleocurrent directions in siltstone beds of the Farmers Member. Also shown are the total number of silstone beds in the unit.

pinchout of the bed and, conversely, wherever this particular bed is found, its base is always mineralized. The thickness of the marcasite mineralization ranges from 0.7 to 2.0 cm (Figure 5), but this thickness variation is unrelated to the thickness of the siltstone, being apparently random areally. The base of the marcasite commonly shows load casting, but no flow marks were found.

Texturally, the marcasite can be divided into two zones. At the base is usually found 1 to 5 mm of pure marcasite. The bulk of the sulfide then occurs above this zone as a replacement of the quartz grains in the turbidite (Figure 6). The quartz is highly corroded, and the grains seem to float in the sulfide. There is no evidence of any size grading or other sedimentary structures involving the sulfide.

Mineralogically the sulfide is almost all marcasite. In contrast,

Table 1. Subunits of the Henley Bed in the Vicinity of Morehead, Kentucky.

Unit	Thickness (cm)		Organic Carbon Content (%)	Number of Samples
3 (top)	20-30	Yellowish gray (5Y, 7/2) to gray (5G, 6/1)	.03 - .06	3
2	45-120	Olive gray (5Y, 4/1) to dark gray (N3)	.20 - .50	6
1	15-90	Greenish gray (5GY, 6/1) glauconitic	.03 - .13	3

All units except 3 thicken sharply to the east. Colors are from GSA rock color chart.

the other units in the section in this area contain pyrite as the predominant sulfide (20 samples were studied). In particular, pyrite is extremely abundant in the underlying Bedford, but little marcasite is found.

A similar sulfide mineralization has been reported from western Kentucky by Kepferle (1972, p. 113-114). There the sulfide is again found in a turbidite unit within the Borden, the Kenwood Siltstone Member. Its sedimentology appears to be very similar to the Farmers (Kepferle, 1977). However, the underlying shale unit, part of the New Providence Shale Member is 36 meters thick at the best-exposed locality for the sulfide, compared with only 1 to 13 meters for the Henley in eastern Kentucky. Also, the mineralization is less continuous than in the Morehead outcrops, occurring as large nodules, cementing the lower part of the siltstone. The sulfide was too fine-grained for unequivocal identification of the minerals in the microscope, but X-ray diffraction showed only pyrite. One example was found in which such a nodule was cutting across the flute and groove casts on the base of the bed, suggesting a post-depositional origin of this sulfide. To summarize, this type of mineralization appears to be moderately widespread in these rocks, in both examples occurring at the boundary between a shale and a siltstone unit, and within the siltstone.

DISCUSSION

How did this sulfide layer form? We have identified four possible mechanisms:

(1) the sulfides were transported by turbidity currents and form the base of the bed because of their higher density,

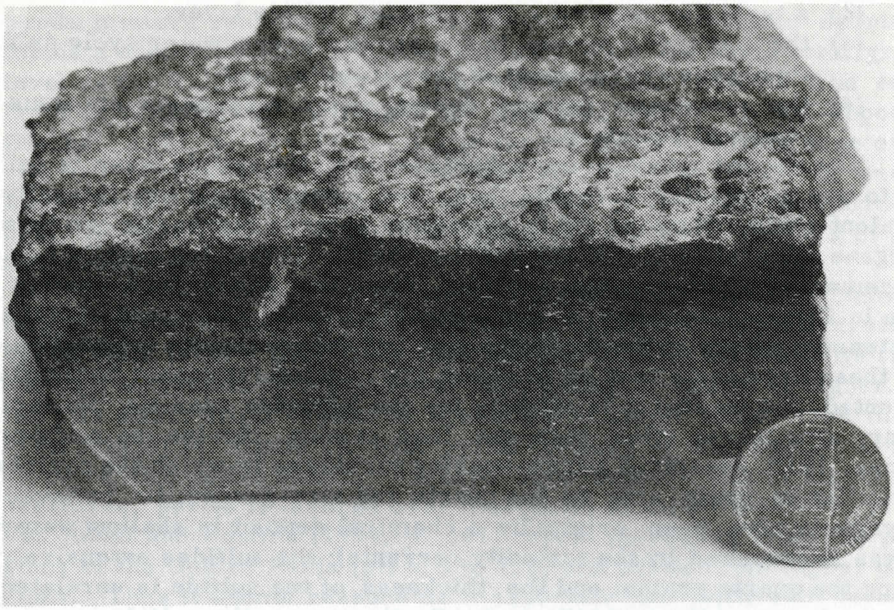

Figure 5. Typical occurrence of marcasite in the Farmers, with base of the bed up. Note that mineralization has proceeded irregularly upward. The load casting found on this bed is more pronounced than on unmineralized beds.

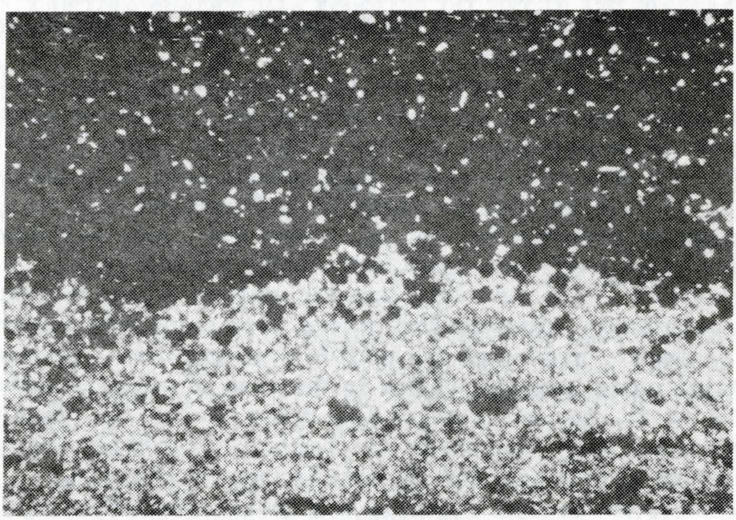

Figure 6. Photomicrograph of marcasite mineralization. The quartz grains are strongly corroded, and replacement seems to stop at a sharp boundary, with scattered marcasite higher. X75.

(2) a turbidite covered a pre-existing sulfide layer,

(3) the marcasite formed in the present groundwater cycle as a supergene replacement of the turbidite, and

(4) the marcasite formed during early diagenesis at some chemical boundary.

The second option is improbable because the sulfide layer is confined to one particular turbidite bed, not extending out into laterally equivalent shale. Number (3) is attractive because marcasite can be a supergene mineral (e. g., Field and Lombardi, 1972), but usually as a replacement of a pre-existing sulfide. Also the heavy load casting shown in Figure 5 suggests a depositional origin. Combined with the co-extensive nature of the turbidite bed and the marcasite mineralization, these arguments lead us to reject (3). Option (1) cannot be totally discounted, but we feel it is unlikely for the following reasons. No sedimentary structures, other than the load casting, involve the sulfides (Skinner, 1958, described pyrite-bearing turbidites in which the pyrites are similar in grain size to the quartz and size-graded; he believed the pyrite to have been originally a chemical deposit in shallow water that was redeposited by the turbidity currents); the sulfides extensively replace the quartz grains; and the thickness of the sulfide is unrelated to the thickness of the turbidite bed. Furthermore, the sulfide is confined to the boundary between a shale unit and a siltstone (plus shale) unit. In the sulfide turbidites described by Skinner, and in the pyrite ore bodies of Portugal and Spain, which are thought to be sulfide turbidites related to contemporaneous volcanism (Schermerhorn, 1971; Strauss and Madel, 1973), there are multiple beds of sulfide. Thus we feel that the marcasite mineralization was caused by a diagenetic process acting after the deposition of the turbidite bed, but before the sediment was completely lithified.

Berner (1969) has shown that in sediments with abundant iron, iron sulfide layers can form by migration of Fe^{++} from the underlying sediment upwards to an organic matter-rich layer where reduction of seawater SO_4^{--} by bacteria produces H_2S, which then precipitates the Fe^{++} as FeS. It may be that the turbidite bed covered such an organic layer, although no evidence for one is preserved. Alternatively, the Henley in this area may have been sufficiently organic-rich to provide a supply of both Fe^{++} and dissolved organic matter, which migrated upward to the permeable siltstone bed where they came into contact with SO_4^{--} from seawater, producing the sulfide precipitation. Certainly the Henley contains more organic matter in the area of mineralization than it does to the north, but not a great deal more (Table 1). Thus we can only say that some such diagenetic mechanism is probably responsible for the formation of this bed, but its details remain obscure.

Another question that arises is why marcasite rather than pyrite. At least one other sulfide layer in rocks of this age is marcasitic: the so-called Leicester Marcasite bed associated with the Tully Limestone (Heckel, 1973, p. 58-60). The sulfide layer in this case is a marcasite-pyrite mixture replacing shell fragments and phosphatic debris. It forms a discontinuous horizon at approximately the same stratigraphic level as the Tully, but beyond its western-most extent. The sulfide layer is believed to represent an extended period of non-deposition (Heckel, 1973, p. 116-117). A similar but more localized zone of pyrite and marcasite is found at the base of the Ohio Shale in eastern Kentucky and southern Ohio. It is made up of abundant phosphatic debris, rounded quartz grains, and approximately equal amounts of pyrite and marcasite.

The only detailed work on marcasite synthesis at low temperature is that of Rickard (1969a, 1969b) who found that both marcasite and pyrite form from a precursor iron sulfide, usually mackinawite (FeS), and that marcasite forms at lower pH, below 7, pyrite at higher. He felt this effect might depend on speciation of sulfur rather than directly on pH (1969b, figure 2), with pyrite forming by the reaction of mackinawite with native sulfur. Rickard also reports that at higher temperatures, pyrite is favored over marcasite at all pH's (1969b, p. 86). Van Andel and Postma (1954, p. 101) report marcasite from low pH regions of the Gulf of Paria (pH 6.5 to 7.0) and pyrite from higher pH regions. Thus it seems reasonable to associate marcasite with deposition at pH$<$ 7. However, such a low value seems unreasonable for the normal marine rocks in this section (pH $\simeq$ 8), so that some other factor may be involved. The other marcasite occurrences in the section appear at surfaces of disconformity or as rims around earlier pyrite nodules, suggesting that marcasite may form when the rate of sulfide growth is low.

The free energy data reported for pyrite and marcasite in Grønwald and Westrum (1976) show that marcasite is everywhere metastable with respect to pyrite; thus its presence in sedimentary rocks suggests that under certain circumstances it must form more quickly than pyrite. This conclusion is supported by the observation that it forms more readily than pyrite at lower temperatures. M. Goldhaber (personal communication) suggests that if the solution from which the sulfide minerals are forming is under-saturated with respect to mackinawite, but still saturated with respect to FeS_2, marcasite may form directly without an FeS precursor (Goldhaber and Kaplan, 1974, p. 595-644, discuss direct formation of FeS_2).

Thermodynamic calculations (Figure 7) show that there is a region, at lower pH and more oxidizing Eh, where FeS_2 is stable, but where FeS cannot form. With no competition from mackinawite, FeS_2 may be able to form directly, although very slowly. If so, its habit should be larger single crystals rather than the distinct spherical aggregates (framboids) that form when the reaction proceeds via a precursor such as mackinawite (Sweeney and Kaplan, 1973). Figure 7 suggests that marcasite may be favored at low pH or under mildly reducing conditions. The latter may apply to the marcasite in the Farmers siltstone, because there is little associated organic matter, and the

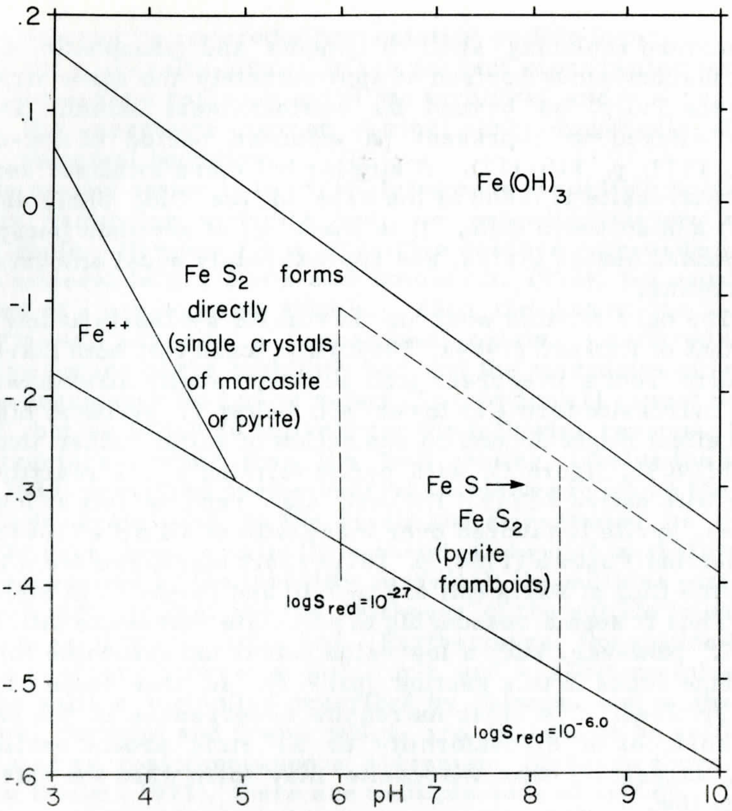

Figure 7. Eh-pH diagram for iron-sulfur at 25°C. $Fe(OH)_3$ is used instead of crystalline hematite to represent the oxidized phase. Note the separate field for FeS within the area of FeS_2 stability. Within this area, FeS forms first. Data from Berner (1971), Grønwald and Westrum (1976).

mineralization is found between somewhat reducing sediments below and more oxidizing above.

For marcasite within the black shale units, this explanation is less likely. Instead note that the FeS field shrinks considerably when the amount of reduced sulfur in the system decreases. In slowly deposited sediments, the rate of sulfate reduction, and consequently the amount of reduced sulfide, decreases substantially (Goldhaber and Kaplan, 1975; Goldhaber and Kaplan, 1974, p. 610). Thus with very low sedimentation rates, mackinawite cannot form, and marcasite may appear instead.

CONCLUSIONS

Prodelta turbidites in the Mississippian clastic sequence of Kentucky contain sulfide mineralization in at least two localities. In both cases the mineralization is found at the boundary between a shale unit and an overlying siltstone unit, within the base of the first siltstone. The sulfide may be localized in the siltstone rather than the shale because of its higher porosity and permeability. In the Morehead, Kentucky, area, the mineralization is almost entirely marcasite with minor pyrite. The mineralization around Morehead is confined to a single turbidite bed, and is absent to the north. This localization may be caused by higher organic matter in the underlying shale.

We feel the marcasite formed during early diagenesis, after the deposition of the turbidite bed, but before complete lithification of the underlying mud. By analogy with other marcasite occurrences in the Devonian of the Appalachian Basin, the predominance of marcasite over pyrite may be controlled by the rate of sulfide growth, slower rates favoring marcasites.

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